KNOWLEDGE GAPS FROM THE IPCC SPECIAL REPORT ON THE OCEAN AND CRYOSPHERE IN A CHANGING CLIMATE AND RECENT ADVANCES

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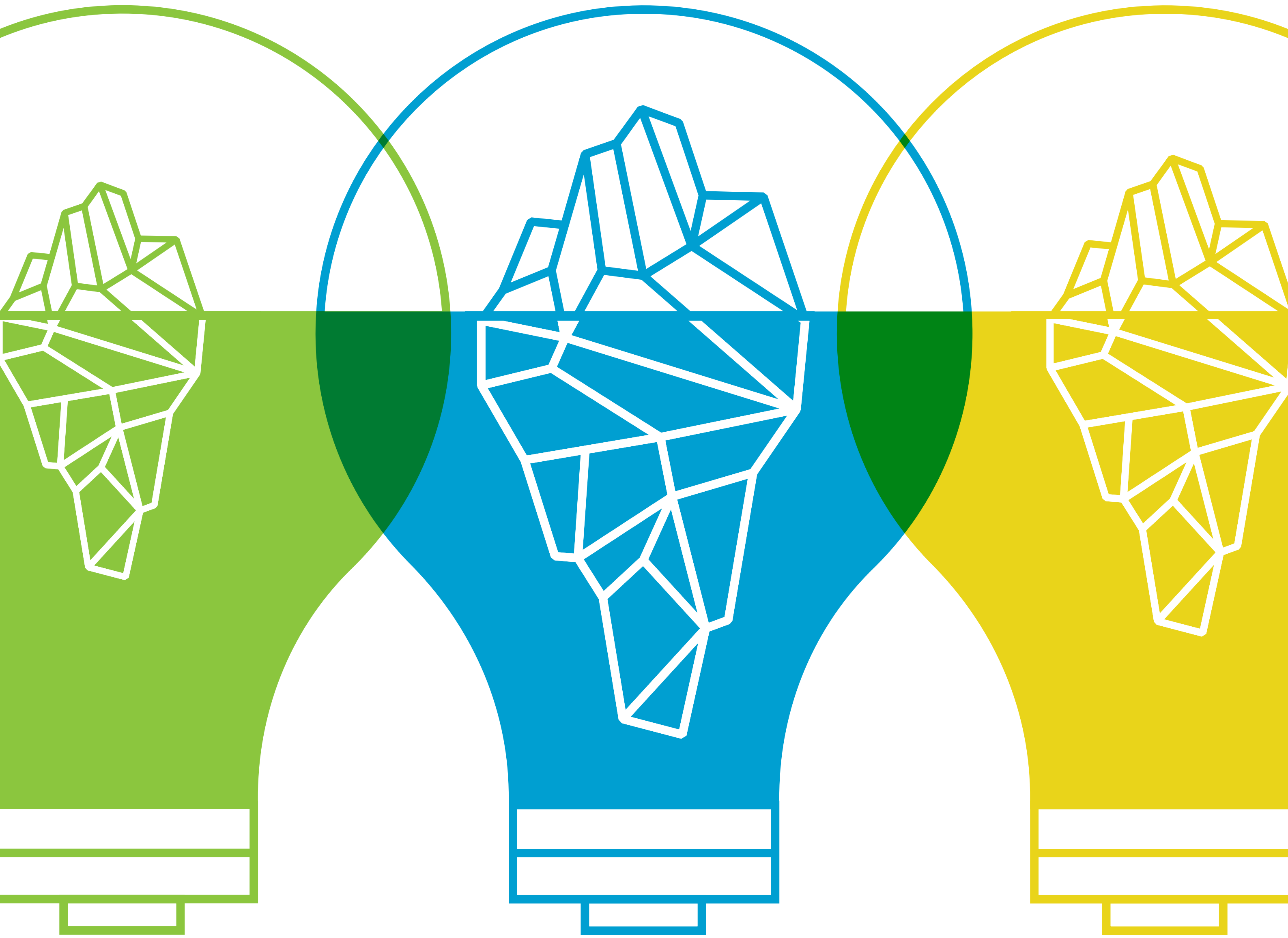

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KNOWLEDGE GAPS FROM THE IPCC SPECIAL REPORT ON THE OCEAN AND CRYOSPHERE IN A CHANGING CLIMATE AND RECENT ADVANCES

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PERSPECTIVE
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Global Data Gaps in Our Knowledge of the Terrestrial Cryosphere

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The IPCC Special Report on Oceans and Cryosphere in a Changing Climate identified major gaps in our knowledge of snow and glacier ice in the terrestrial cryosphere. These gaps are limiting our ability to predict the future of the energy and water balance of the Earth's surface, which in turn affect regional climate, biodiversity and biomass, the freezing and thawing of permafrost, the seasonal supply of water for one sixth of the global population, the rate of global sea level rise and the risk of riverine and coastal flooding. Snow and ice are highly susceptible to climate change but although their spatial extents are routinely monitored, the fundamental property of their water content is remarkably poorly observed. Specifically, there is a profound lack of basic but problematic observations of the amount of water supplied by snowfall and of the volume of water stored in glaciers. As a result, the climatological precipitation of the mountain cryosphere is, for example, biassed low by 50–100%, and biases in the volume of glacier ice are unknown but are likely to be large. More and better basic observations of snow and ice water content are urgently needed to constrain climate models of the cryosphere, and this requires a transformation in the capabilities of snow-monitoring and glacier-surveying instruments. I describe new solutions to this long-standing problem that if deployed widely could achieve this transformation.

Keywords: snowfall, glacier, ice, water, SWE, survey, instrument

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INTRODUCTION

The IPCC Special Report on Oceans and Cryosphere in a Changing Climate (SROCC) highlighted key strengths and weaknesses in our understanding of the cryosphere (IPCC, 2019). It showed that we can observe with *high* or *very high* confidence (IPCC confidence definitions in italics) loss of mass from the ice sheets (SROCC section A.1.1), declining sea ice (SROCC A.1.4), and snow cover (SROCC A.1.2), and can produce probabilistic climate projections across a range of future scenarios. For many aspects of the cryosphere, however, it is only the trajectories of change that can be projected with *medium* to *very high* confidence. Their magnitudes are either not quantified, have large uncertainties or are quantified with *low* confidence. These include terrestrial snow cover (SROCC A.1, A.7.7, A.4.1, A.1.4, A.1.2, B.1.3), mountain and Arctic water resources (SROCC A.7.6, B.1.6, B.7, B.4.3), permafrost thaw (SROCC B.7.2), mountain and polar species distribution, biodiversity and biomass (SROCC B.4, B.4.1, B.4.2) and disaster risk (SROCC B.1.5, B.7.1).

Major deficiencies in snow and ice observations from mountain ranges and the Polar Regions contribute to this uncertainty. "Clear knowledge gaps" exist in current glacier-ice volumes and the spatial and temporal variation of snow cover (SROCC section 2.5). Time-series of snow water equivalent (SWE) show "reasonable consistency" when averaged by continent but considerable disagreement in spatial pattern (SROCC section 3.4). Knowledge of SWE trends is "inadequate" (SROCC section 3.7), and mountain precipitation trends globally are "highly uncertain" due to large natural variability and "intrinsic uncertainties" in measurements (SROCC section 2.5). Long-term observations are particularly scarce in High Mountain Asia (HMA), Northern Asia and South America (SROCC section 2.2.2). Similarly over Arctic land, precipitation measurements are "sparse and highly uncertain" (SROCC section 3.7). Atmospheric re-analyses suggest a recent Arctic-precipitation increase but the wide model spread gives only *low confidence* in reanalysis-based closure of the Arctic freshwater budget. A declining trend in snow-depth in the Russian Arctic was assigned *medium confidence* as the "pointwise nature" of weather-station measurements does not capture prevailing conditions across the landscape. A shift in the timing of maximum snow depth was detected for the North American Arctic but no comparable analysis is available for Eurasia (SROCC section 3.7).

Snow and ice strongly modify the albedo and insulation of land and sea surfaces, the wetting or drying (and greening or browning) of the terrestrial Arctic, and the mass balance of all of the world's glaciers and ice sheets. These weaknesses therefore critically impact our understanding of cold-region water and energy balances, with global consequences. Over the recent past, global glacier mass loss was as great as that from the Greenland Ice Sheet (the single largest source of sea level rise), but with 10-fold greater uncertainty (**Figure 1**). Due to the limited number of well-observed glaciers, there is only *medium confidence* in the ability of glacier models to reconstruct past sea-level change (SROCC section 4.2.2.2.3). Climate models also fail to reproduce a pre-1970 Greenland warming and resulting sea-level rise, hence there is only *medium confidence* in the ability of these models to predict future glacier and ice sheet surface mass balance (SMB) (SROCC section 4.2.2.6). These issues are reflected in projected future glacier losses under RCP2.6 by end-of-century that have a "likely" range of uncertainty of ±40%, using models calibrated with only "limited observations" and "diverging initial glacier volumes" (IPCC, 2019, CCB.6).

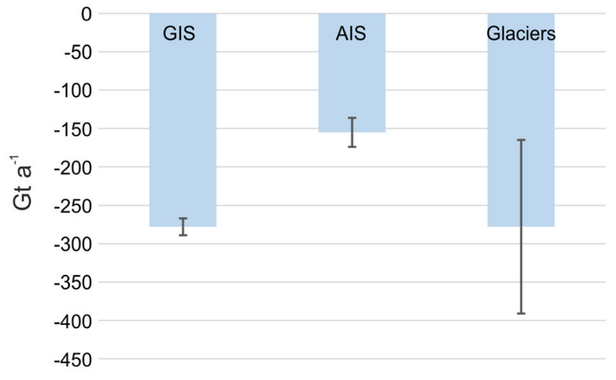

FIGURE 1 | Annual rates of ice loss for 2006–2015 from the Greenland Ice Sheet (GIS) and Antarctic Ice Sheet (AIS) (IPCC, 2019, Table 3.3) and all other global glaciers (IPCC, 2019, section 2.2.3, Table 2.A.1). The glacier losses are as great as Greenland's but with much greater uncertainty.

KEY OBSERVATION GAPS IN THE CRYOSPHERE

Snowfall

Snowfall seasonally covers a third of all land (NSIDC, 2020) and exceeds 3000 Gt of transient water storage (Pulliainen et al., 2020) but snowfall SWE remains difficult to measure, particularly over mountains and the Polar Regions. Globally, most observations come from weather stations such as those contributing to the Global Historical Climate Network (GHCN), a quality-controlled database of 100,000 daily measurements (Menne et al., 2012) of which 26% report precipitation in the cryosphere and only 6% in the mountain cryosphere (**Figure 2A**). Only a single active GHCN station on the dry Tibetan Plateau reports daily precipitation in the combined 566,000 km^2 Himalayan headwaters of the Brahmaputra, Indus and Ganges river basins above 4,000 m altitude (**Figure 2A**). Furthermore, the number of weather stations has decreased and is now at its lowest in over 100 years (Fick and Hijmans, 2017).

These observations do not adequately represent even the climatological-average precipitation in these environments. Gridded precipitation climatologies interpolated from GHCN station data [e.g., the "WorldClim v2" 1970–2000 mean (Fick and Hijmans, 2017)] correlate reasonably well (0.86) on the global scale with test data but universally less well in mountain ranges. More significantly, they systematically underestimate precipitation in much of the cryosphere by 50–100% (**Figure 2B**). WorldClim v2 precipitation in the cryosphere requires an average bias-correction factor (inferred from streamflow data) of 1.52 vs. 1.05 for the rest of the world (Beck et al., 2020). In the mountain cryosphere the global average bias-correction factor is 1.46 for annual precipitation, and 1.61 in winter. Regionally the factor is 1.55 annually in the HMA (1.90 in winter and 1.44 in summer), 1.5–2 annually throughout the Arctic, and up to 2.05 for the Andes in winter (**Figure 2B**). Such biases among mountains are also present in comparable gridded precipitation products produced by various methods, including PREC/L (Chen et al., 2002), CHELSA V1.2, CHPclim V1, GPCC V2015, GPCP V2.3, and MERRA-2 (Beck et al., 2020), and regional hydrological assessments (e.g., Wortmann et al., 2018; Pritchard, 2019). Although these bias corrections have been calculated, they apply only to climatological averages and not the daily timescales of weather models, and also have considerable uncertainty due to the challenges involved in calibrating precipitation from streamflow: the range in possible correction factors in the Andes locally exceeds 4.00 (**Figure 2C**).

In part these precipitation biases (that are worst in winter and in mountains) reflect a spatial-sampling problem in

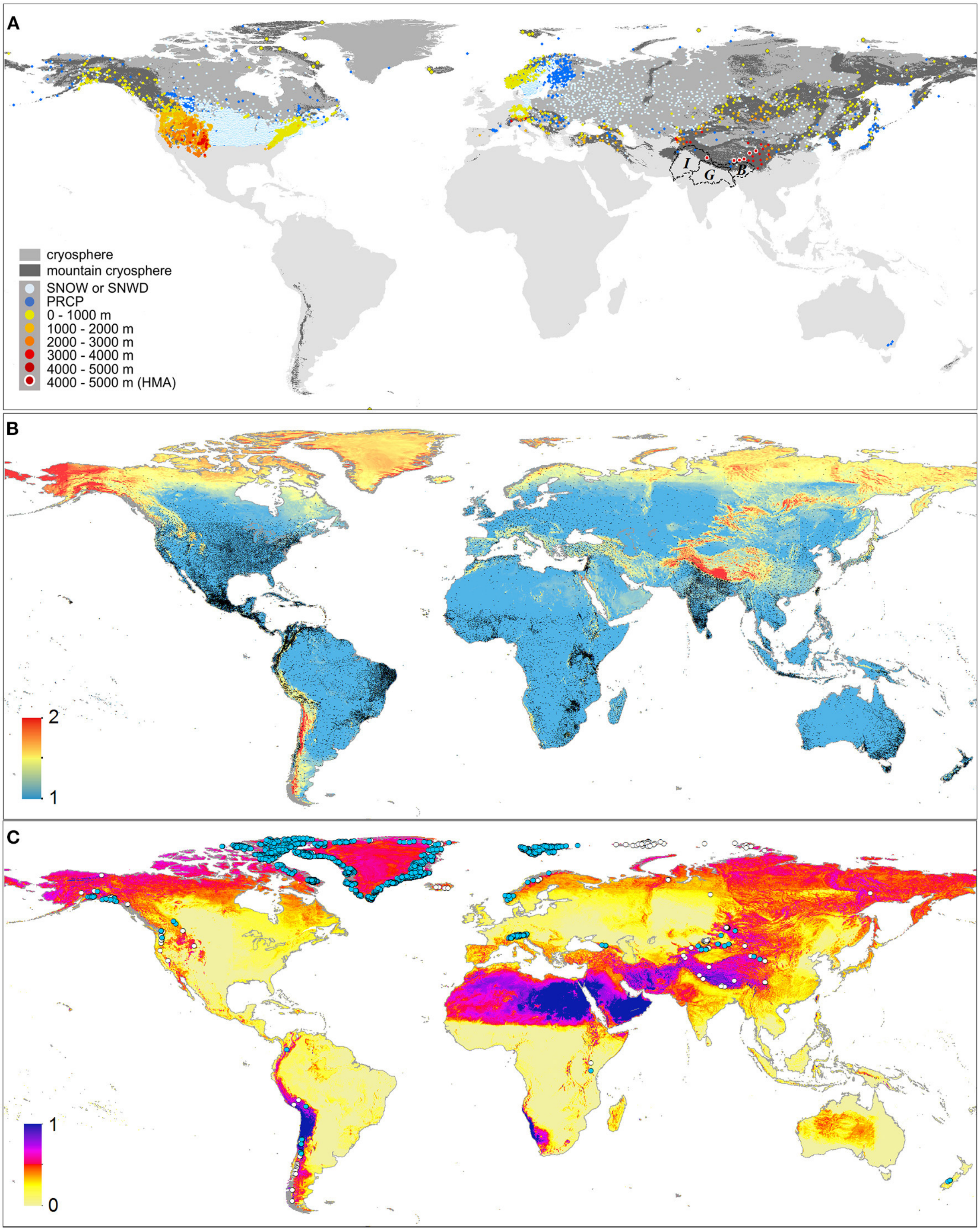

FIGURE 2 | (A) Weather stations (coloured dots) in the Global Historical Climate Network that lie within the terrestrial cryosphere (mid grey), or more specifically within the mountain cryosphere (dark grey) and whose daily observations are ongoing (in 2020/21). Light blue dots show stations in the lowland terrestrial cryosphere that report either snowfall water content (SNOW) or depth (SNWD). Dark blue dots show stations that reported precipitation (PRCP) but do not distinguish snow observations. Yellow to red dots show stations within the mountain cryosphere that report SNOW or SNWD, coloured according to altitude. Black polygons show the river basins of the Indus (*I*), Ganges (*G*), and Brahmaputra (*B*). The cryosphere is defined as having a mean monthly temperature <0°C in January and/or July. The mountain cryosphere is the intersection between these areas and mountain areas (Karagulle et al., 2017). This highlights the paucity of snow observations in much of the mountain and Arctic cryosphere. **(B)** Bias-correction factors (blue-to-red scale, after Beck et al., 2020) that must be applied to the observation-based "WorldClim v2" annual climatological precipitation product (Fick and Hijmans, 2017) in order to agree with observations and models of catchment hydrology. All WorldClim v2 weather stations that reported precipitation (RAINID) in the 1970–2000 period are shown by black dots. **(C)** The uncertainty range in the estimated bias correction factors in 2b (the upper bound of the factor minus the lower bound) (after Beck et al., 2020), and glaciers with a reported average thickness only (white dots) or thickness-profile data (blue dots) (GlaThiDa Consortium, 2020, accessed February 2021).

the cryosphere, but they also reflect a problem of inherent measurement bias. Mountain precipitation, and particularly snowfall, is relatively heterogeneous (Gerber et al., 2018) and so is often not well-sampled regionally by the cryosphere's sparse station network or locally by the very small physical size of pluviometers (<0.3 m diameter) relative to local snowfall variability or to the kilometre-scale grid cells of precipitation products (e.g., Dozier et al., 2016; McCrary et al., 2017; Sturm et al., 2017; Yao et al., 2018; Haberkorn, 2019; Yoon et al., 2019). Snowpack SWE was found to have a standard deviation of 21% within a plot of only 20 × 8 m, for example (Haberkorn, 2019). These sampling problems on the large and small scale are compounded by the characteristic snow "undercatch" of pluviometers that leads to a precipitation low-bias of up to 90% in windy conditions (e.g., Burgess et al., 2010; Beck et al., 2020).

Pluviometer undercatch can be partially mitigated by fences and baffles, but these are rarely in operational use (Yang, 2014). Other daily operational measurements of snowpack SWE (e.g., snow pillows, scales) or SWE proxies (e.g., sonic rangers, radiometers) also observe only small areas (Dozier et al., 2016), and scaling up from the snow-pillow scale to grid scale, for example, was found to introduce a bias of up to 200% (Molotch and Bales, 2005). Together, it is these sparse and biassed snowfall and snowpack measurements that form the main foundation for developing, calibrating and validating physical weather and climate models of the cryosphere and this inevitably limits the confidence level of their predictions (e.g., IPCC, 2019).

Case Study: Greenland Sea-Level Contribution

Up until the 1990s, the Greenland Ice Sheet's annual ∼700 Gt of snow accumulation was approximately balanced by losses of ∼40 Gt to sublimation, ∼260 Gt to surface melt runoff, and ∼410 Gt to iceberg discharge (van den Broeke et al., 2017). Subsequent acceleration of glacier flow and a decrease in SMB led to net losses of >200 Gt per year (**Figure 1**) but with large interannual fluctuations (Shepherd et al., 2020). While our understanding of mass-loss processes has advanced considerably, the snowfall input (hundreds of gigatons larger than each loss term) remains very poorly observed and understood (Hanna et al., 2020). Field observations are mostly infrequent and limited to the dry ice-sheet interior, coming from: (a) ∼133 firn cores with at-best annual resolution (Burgess et al., 2010); (b) monthly measurements at Summit Station; or (c) occasional radar surveys for multi-year snow depths (Montgomery et al., 2018). Near the wetter coastal margins, 40 sensors of the PROMICE and GC-NET arrays monitor snow height (Steffen et al., 1996; Citterio et al., 2015; Cappelen, 2018) but only 15 pluviometers make sub-daily measurements of snowfall SWE comparable to precipitation in weather models (Cappelen, 2018). Up to 90% of SWE reported by these sensors actually consists of estimated undercatch corrections (Yang et al., 1999; Bales et al., 2009), and their combined area of <1 m^2 is likely representative of <1 km^2 of Greenland's 39,000 km-long coastline (van den Broeke et al., 2017).

Greenland's snowfall inputs are likely to have changed, being sensitive to the same trends in the North Atlantic Oscillation and cyclogenesis that have affected its surface melt rate (Rogers et al., 2004; Bevis et al., 2019). Furthermore, snowfall is not only a mass source but also influences melt losses through its control of albedo and surface roughness, its thermal mass, its properties as an insulator and its role as a meltwater aquifer (van den Broeke et al., 2017; Ryan et al., 2019). Our poor understanding of snowfall is highlighted by the difficulties that regional climate models have in reproducing Greenland's seasonal snowline (Ryan et al., 2019) and in significant disagreements in historical accumulation and SMB (Fettweis et al., 2020; Hanna et al., 2020). Mass losses for 2003–2012 retrospectively calculated by thirteen models ranged from 1066 to 6034 Gt, with large ensemble uncertainty of ±1253 Gt (±48%, or ±3.5 mm of sea level) and local uncertainty of up to 2 m W.E. per year (Fettweis et al., 2020). This wide spread was due only to differences in SMB (dynamic losses were standardised), demonstrating the substantial lack of model consensus on snowfall-dominated surface processes (Bales et al., 2009; van den Broeke et al., 2017; Montgomery et al., 2018; Fettweis et al., 2020).

The model *uncertainty* range for Greenland's recent SMB equates to 1 mm of global sea level over 4 years (Table 4.1 in IPCC, 2019), and under the three main climate (RCP) scenarios, the SMB uncertainty range in predicted contributions by 2100 is 6 cm, 7 cm or 14 cm (Table 4.4 in IPCC, 2019), amounting to 50–70% of Greenland's total mass-loss uncertainty (Aschwanden et al., 2019). Similar uncertainties persist in newer CMIP6 model runs under the SSP forcing scenarios (Hofer et al., 2020). The magnitude of these uncertainties in SMB and sea-level rise imply an uncertainty in annual coastal-flooding costs to the global economy by 2100 of around $1–2 trillion (Jevrejeva et al., 2018).

Glacier Thickness

The thickness of mountain glaciers is much less well-surveyed than that of the ice sheets (Pritchard, 2014) and projections of future mass-loss are highly sensitive to the initial ice volume (Hanna et al., 2020). Some thickness information is available from 5,141 glaciers (2.6%) globally (**Figure 2C**) (Welty et al., 2020, accessed 25 February 2021) but is poorly-distributed, sampling only 0.07% of 95,000 HMA glaciers, for example (RGI Consortium, 2017), and has questionable accuracy. A "mean thickness" is reported for 67 small (mean 2.7 km^2) and one large glacier (Fedchenko, 824 km^2) but only 8 glaciers report survey profiles, all of which are short (<3 km), thin (mean 60 m), clustered in the northern ranges and largely predate accurate GPS survey-control (mean year 1987). This shortage of thickness measurements reflects the difficulties of surveying, particularly on remote, high, debris-covered glaciers (Pritchard et al., 2020).

Given the data scarcity, glacier volumes are estimated from scaling relationships with area and inversions of local thickness from surface characteristics (topography and SMB) using principles of ice-flow constrained with measurements where available (e.g., Langhammer et al., 2019a). Estimates vary widely however: a 5-model ensemble recently revised downwards by 46% the estimated HMA ice-volume, though each model frequently produced local deviations of up to twice the observed mean thickness (Farinotti et al., 2019). Globally, the ensemble-average uncertainty is estimated to be ±26% (Farinotti et al., 2019) but without representative observations the absolute accuracy is unknowable for most mountain ranges.

The thickness-inversion approach (Huss and Farinotti, 2012) was tested on Austrian glaciers with extensive topographic and climatic data, and unusually widespread, well-distributed thickness measurements for 58 glaciers representing >40% of Austria's glacierised area (Helfricht et al., 2019). Key to the inversion process is a mass-balance-gradient parameter that must be calibrated with SMB and thickness observations. When carefully optimised to individual glaciers within the data-rich Austrian subset, this key parameter was found to exhibit a large spatial and temporal spread and yielded a thickness uncertainty of 25–31%, with 5% residual bias. Without such tuning the bias was +25% (Helfricht et al., 2019) though, highlighting the importance of extensive ice-thickness and SMB calibration measurements even locally within a mountain range.

The thickness of non-flowing, stagnant areas of glacier ice cannot be calculated from scaling or inversion, and these are increasingly common on debris-covered glaciers as their surfaces lower and flatten. A minimum slope threshold of 2° was applied to an inversion of HMA glaciers to exclude such areas (Kraaijenbrink et al., 2017), but 12% of glaciers have slopes <2° and these tend to be disproportionately large with thickness ∼3 times greater than the regional average (ICIMOD, 2011). At least 1,250 km^2 (4%) of the regional glacier ablation area is effectively stagnant [flow rate <5 m a^{-1} (Kraaijenbrink et al., 2017; after Dehecq et al., 2019)], an area likely containing more than 100 km^3 of ice in just this region.

Case Study: Nepal Glacier Lifespan

Recent field surveys with a low-frequency ice-penetrating radar have produced detailed profiles of thickness for the slow-flowing, debris-covered lower tongues of three Nepal glaciers, allowing modelled ice thicknesses to be tested. At these profiles the modelled thickness (Kraaijenbrink et al., 2017) for Ngozumpa, Nepal's largest glacier, was biassed by −32% (modelled 184 m, measured 270 m), and for Lirung and Langtang glaciers, the biases were −77 and +31%, respectively (Pritchard et al., 2020). The significance of these biases is clear from the glaciers' projected lifespans based on recent thinning rates. For Ngozumpa, this is reduced from around 420 to 290 years. For Lirung and Langtang, projections based on measured thickness are around 300 and 200 years, respectively vs. 70 and 260 years from modelled thickness.

RECENT ADVANCES IN MEASURING SNOWFALL AND GLACIER THICKNESS

To address the sampling problems and measurement biases of existing snowfall instruments, new spatially-integrated methods have been developed based on monitoring sub-surface water pressure as it responds to surface snow-loading. The "geolysimeter" approach employs a sensor in an aquifer borehole to monitor changes in groundwater pressure, but is limited in potential application to suitably confined aquifers and by the cost of borehole drilling (Smith et al., 2017). A more widely-applicable approach is based on monitoring winter water pressure in lakes as it responds instantaneously to the mass of precipitation falling onto the lake surface (Pritchard et al., 2021). Importantly, both methods help eliminate bias in the calibration and validation of weather models as they sense on hourly timescales specifically the water equivalent of snowfall, avoid undercatch as the sensors are submerged, and average over large areas (e.g., several square kilometres). The lake method has been used in alpine and Arctic lakes that were 1 million to 274 million times larger than the nearest available conventional pluviometers, and through 25 snowfalls over a winter at a Swiss mountain lake, average uncertainty in the snowfall rate was calculated as ±0.1 mm W.E. h^{-1} (Pritchard et al., 2021).

While the latter method is limited to lake sites, these are abundant in the world's mountain ranges and glacier margins. The World Meteorological Organisation recommends a precipitation sampling-density of 0.4 pluviometers per 100 km^2 among mountains, with each observing an area of ∼0.05 m^2 (WMO, 2018; Haberkorn, 2019). Within just the HMA mountain cryosphere there are over 25,000 lakes at altitudes ranging from 1700 to 6200 m (mean 4,710 m) and covering a total of 1,735 km^2 (Wang et al., 2020). This gives a potential sampling density of 0.7 lakes per 100 km^2 and an observable area of 48,000 m^2 per 100 km^2, much larger than possible with pluviometers. Globally there are also over 14,000 lakes situated on mountain glaciers, covering 8,950 km^2 (Shugar et al., 2020). Coastal Greenland has 3,347 lakes within 1 km of the ice sheet margin totaling ∼3,000 km^2 (at 4 lakes or 3,600,000 m^2 observable area per 100 km^2) (How et al., 2021). Instrumenting a small subset of these many lakes could narrow the range of predicted Greenland SMB, and hence the future rate of sea level rise, by allowing unrealistic ensemble model outputs to be culled and by providing calibration of the climate-model physics needed to improve their predictions.

For ice-thickness surveying, recent progress has also been made in the development of helicopter-borne ice-sounding radars in the European Alps (Rutishauser et al., 2016; Langhammer et al., 2019b) and challenging, large, debris-covered Himalayan glaciers (Pritchard et al., 2020). Being modular, lightweight and capable of low frequencies, the latter system is particularly suited to reach otherwise inaccessible glaciers. It has, for example, been used over heavily-crevassed Arctic tidewater margins (Pritchard et al., 2020) and the major glaciers of the Everest area through ice >200 m thick and at altitudes up to 6,500 m (https://www.bas.ac.uk/project/bedmap-himalayas/). Measurements from an enlarged sample of such glaciers would improve model calibration and de-biassing of thickness-inversions in many more ranges beyond the European Alps.

CONCLUSIONS AND FUTURE PRIORITIES

Seasonal snowfall and glacier ice play important ecological, hydrological, socio-economic, and climatic roles within the Earth system, but IPCC SROCC identified large biases and uncertainties in the present-day magnitude of these major water-cycle components and even greater uncertainty in their future evolution. These uncertainties are globally significant, producing for example large ranges in possible sea level rise and coastal flood risk over coming decades, and large ranges in the potential lifespan of the water supply from mountain glaciers. A primary cause of these biases and uncertainties is a lack of basic measurements in the cryosphere, reflecting the practical difficulties of monitoring and surveying in such environments. New instruments have recently been developed to overcome these difficulties, and a key priority now is to deploy them widely to collect representative observations of snowfall and ice-thickness that are sufficient to constrain weather, climate and glacier models. This constraint should empower the models to predict the cryosphere's future with greater confidence.

DATA AVAILABILITY STATEMENT

Publicly available datasets were analysed in this study. This data can be found at: GlaThiDa: https://www.gtn-g.ch/data_catalogue_glathida/; WorldClim v2: https://www.worldclim.org/; Bias corrections: http://www.gloh2o.org/pbcor/; GHCN: https://www1.ncdc.noaa.gov/pub/data/ghcn/daily/; Mountains: http://rmgsc.cr.usgs.gov/outgoing/ecosystems/Global/; PREC/L precipitation data: https://psl.noaa.gov/data/gridded/data.precl.html.

AUTHOR CONTRIBUTIONS

The author confirms being the sole contributor of this work and has approved it for publication.

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Untold Stories: Indigenous Knowledge Beyond the Changing Arctic Cryosphere

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Scientific attention to climate change in the Arctic has spurred extensive research, including many studies of Indigenous knowledge and the effects of climate change on Indigenous peoples. These topics have been reported in many scientific papers, books, and in the IPCC's 2019 Special Report on the Ocean and Cryosphere in a Changing Climate (SROCC), as well as attracting considerable interest in the popular media. We assembled a set of peer-reviewed publications concerning Arctic Indigenous peoples and climate change for the SROCC, to which we have added additional papers discovered through a subsequent literature search. A closer look at the 76 papers in our sample reveals additional emphases on economics, culture, health and mental health, policy and governance, and other topics. While these emphases reflect to some degree the perspectives of the Indigenous peoples involved in the studies, they are also subject to bias from the interests and abilities of the researchers involved, compounded by a lack of comparative research. Our review shows first that climate change does not occur in isolation or even as the primary threat to Indigenous well-being in the Arctic, but the lack of systematic investigation hampers any effort to assess the role of other factors in a comprehensive manner; and second that the common and perhaps prevailing narrative that climate change spells inevitable doom for Arctic Indigenous peoples is contrary to their own narratives of response and resilience. We suggest that there should be a systematic effort in partnership with Indigenous peoples to identify thematic and regional gaps in coverage, supported by targeted funding to fill such gaps. Such an effort may also require recruiting additional researchers with the necessary expertise and providing opportunities for inter-regional information sharing by Arctic Indigenous peoples. As researchers who are visitors to the Arctic, we do not claim that our findings are representative of Indigenous perspectives, only that a more accurate and comprehensive picture of Arctic Indigenous peoples' knowledge of and experiences with climate change is needed. Our analysis also reflects some of the SROCC knowledge gaps and the conclusions provide suggestions for future research.

Keywords: Indigenous, Arctic, climate, adaptation, health, economics, culture, governance

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INTRODUCTION

The Arctic is changing rapidly, spurring much scientific and media attention (e.g., Christensen et al., 2013; Arnold, 2018; NOAA, 2020). The Arctic cryosphere in particular is regarded as one of the most visible signs of global warming, as sea ice retreats, snow decreases, glaciers and ice caps melt, and permafrost thaws (Meredith and Sommerkorn, 2019). A changing cryosphere has far-reaching implications for biology (Wassmann et al., 2020) and society in the Arctic (AMAP, 2017a,b,c) and beyond (Moon et al., 2019).

In addition to standard scientific studies in the Arctic, researchers have documented Indigenous knowledge about the Arctic environment and its changes (e.g., the publications listed in the **Supplemental Material**). Indigenous Peoples have lived in the region for thousands of years, accumulating extensive and detailed understanding of the environment and of human relationships with the lands, waters, air, plants, and animals to be found there (e.g., Watt-Cloutier, 2018). Such information is invaluable for its own sake as well as for the depth of time and breadth of coverage that can be found from no other sources (e.g., Berkes, 2012; Thornton and Bhagwat, 2021). Attention to Indigenous knowledge has been growing, from a dedicated chapter in the *Arctic Climate Impact Assessment* (Huntington and Fox, 2005) to more recent inclusion in the work of the Intergovernmental Panel on Climate Change (Meredith and Sommerkorn, 2019).

The recent IPCC Special Report on the Ocean and Cryosphere in a Changing Climate (SROCC; IPCC, 2019) is a case in point. Martin Sommerkorn, one of the lead authors of the chapter on Polar Regions, contacted us to help his team gather available studies documenting Indigenous knowledge and related information about the cryosphere. The goal was to try to include as much of that information as possible in the chapter. Although we knew that many studies had been done, we did not know at the start how many publications would be available that met the IPCC criteria for inclusion: that the works be published or accepted for publication in the scientific or technical literature by a cut-off date specified for each report (e.g., IPCC, 2018). As described below in Methods, we assembled an annotated list of publications and sent that to the chapter authors for their use.

In assembling and reviewing these publications, we noticed some patterns. First, few if any of the papers stopped at reporting about the components of the cryosphere. They additionally discussed many other changes taking place in the Arctic, or the implications of cryosphere change for other aspects of Arctic communities. Second, the coverage of these additional topics was uneven around the Arctic, suggesting differences in relative importance or differences in the design and focus of the studies that generated the papers. Third, the narrative from many of the papers diverged from the common story of inevitable doom and gloom facing the Arctic and its inhabitants. The changes are serious and pose a major threat, but Arctic communities are also capable of responding to these changes (Huntington et al., 2019).

In this paper, we provide a brief review of the SROCC knowledge gaps and explore all three above-noted observations, based on the initial literature search for the SROCC and a subsequent expansion to include papers published since the SROCC deadline for the present review. Together, our findings suggest that the changing Arctic cryosphere has not been fully explored, either in terms of the understanding of cryospheric changes from the perspective of Arctic Indigenous Peoples, or with regard to the implications that change has for them. The increasing number of publications on Indigenous knowledge and perspectives provides more material reflecting their views, and the inclusion of this information in assessments such as the SROCC is a welcome step. Nonetheless, greater attention is needed across the full range of effects on Arctic communities and across the entire Arctic. In addition, we note that the perspectives of Arctic Indigenous Peoples as reflected in these publications often suggest a degree of hope that is typically missing in visitors' assessments of the prospects for Arctic communities in a changing climate.

METHODS

Positionality

In a paper about Indigenous knowledge in the Arctic, we are obliged to point out that neither author is an Arctic Indigenous person. We are scholars who live outside the Arctic (LEM in Victoria, British Columbia, Canada, and HPH in Eagle River, Alaska, USA), though we both travel to the Arctic frequently for work and more. Our findings and views thus reflect our position as outsiders or visitors, and may not match those of Arctic Peoples. This paper focuses on academic reporting about Indigenous knowledge and views in the Arctic. We feel qualified to speak to academic treatment of the topic, though we recognize that what is reported in academic papers may not reflect the full scope of views of Arctic Indigenous Peoples. We thus conclude the paper by recommending further work to better engage Arctic Peoples, in their own voices, to better include their experiences, conclusions, and recommendations (Pfeifer, 2018).

SROCC Preparation

We were asked to help find sources of Indigenous and traditional knowledge relevant to the Polar Regions chapter of the SROCC (Meredith and Sommerkorn, 2019). No resources were available to support our work, but we decided that we could nonetheless make a useful contribution within the limits of the time and effort we could volunteer. We conducted a chain-referral search, starting with colleagues whom we knew to be active in this field. We asked for any papers meeting the IPCC criteria: peer-reviewed papers or books, published between 2010 and 2017. We also asked our colleagues to forward our request to others who might be able to help. In addition, we conducted a search on Google Scholar. In the end, we received 110 papers from many colleagues, active in the field in all Arctic countries. We reviewed the 110 papers and selected 56 that met the IPCC criteria and were on topic for the Arctic cryosphere, in that they included traditional knowledge about or Indigenous views on snow, sea ice, permafrost, and the implications of change for Arctic Peoples and communities. A few papers published prior to 2010 were included if the material was particularly pertinent. We added our own notes about each of the papers, including selected quotes that seemed particularly pertinent to the SROCC chapter, and sent the annotated list to the chapter's authors.

While conducting this exercise we realized that to varying degrees the papers reflected Indigenous People's views related not only to climate but also to other environmental, social, economic, political, health, mental health, and cultural factors affecting them.

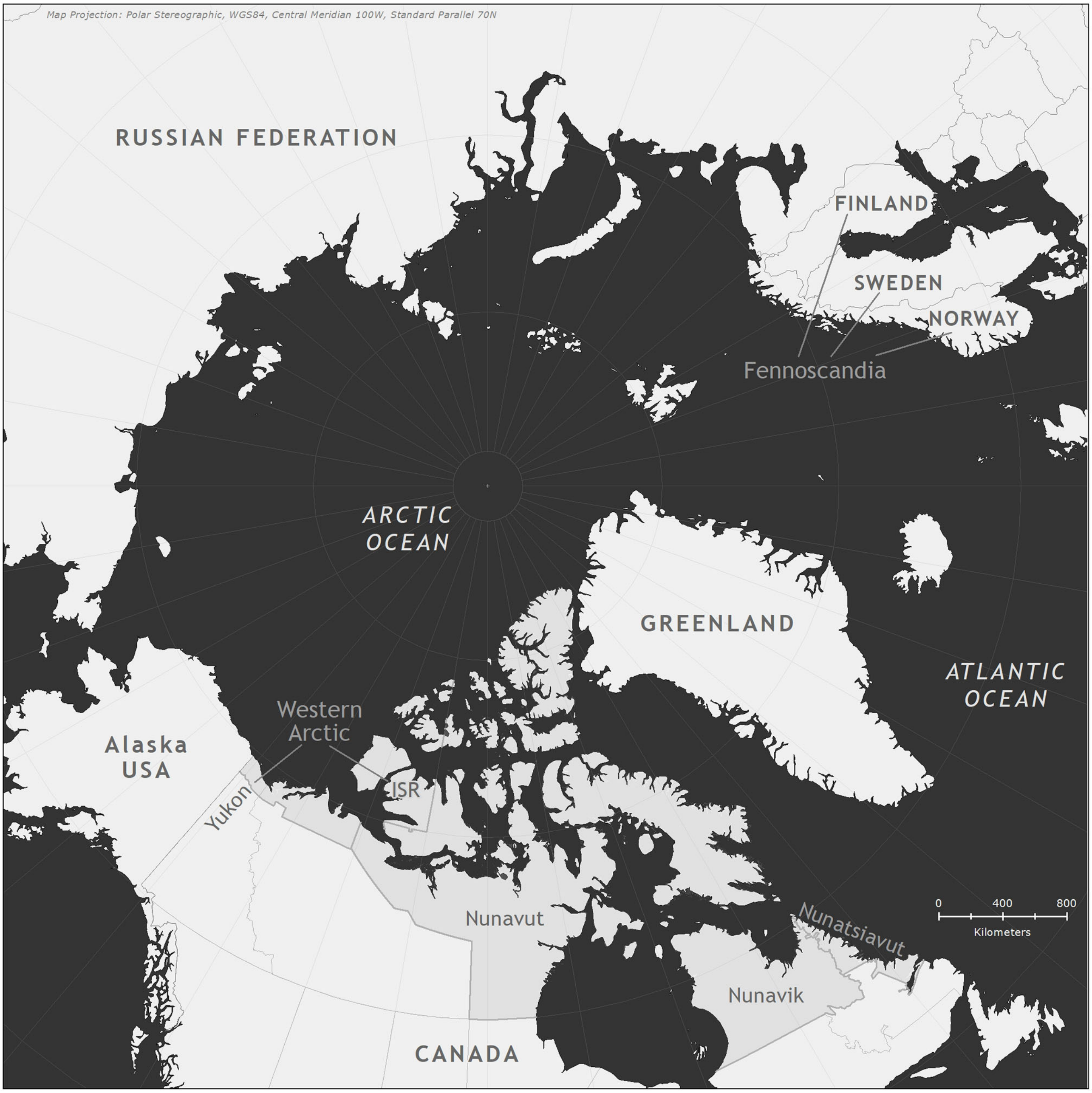

FIGURE 1 | Map of the Arctic showing countries and regions considered in this paper. ISR, Inuvialuit Settlement Region, included in "Western Arctic" for our purposes.

Literature Search

Noting that in some cases the papers we compiled for the SROCC chapter reflected some of the interconnectedness that Indigenous people mention when they share their traditional knowledge, we decided to expand the initial search and update the initial list of papers to include what else had been published since 2017. In one way, the literature search on its own might have been sufficient to create the sample of papers used here, since most or all of the papers on our SROCC list were also found during the literature search. In another way, though, the SROCC list was important because it started us thinking about the additional topics we consider here. In our SROCC work, we had seen a clear difference from country to country in terms of involvement of Indigenous people in biophysical research. We thus targeted the subsequent search by country. Because of the relative sparsity of such studies in the European Arctic region, we searched for Fennoscandia (i.e., Norway, Sweden, and Finland) as a whole (**Figure 1**).

The focus of SROCC Chapter 3: Polar Regions is the ocean and the cryosphere. In dealing with Indigenous Peoples, the focus was primarily on lowland and coastal communities, regarded as the most exposed and vulnerable to cryospheric changes (Meredith and Sommerkorn, 2019). Thus, the main Indigenous groups we considered in our search were coastal and lowland communities, which are predominantly the various Inuit groups in Alaska, Canada, and Greenland. In the case of Russia and Fennoscandia, there are not as many Indigenous coastal communities affected by changes in the cryosphere. In those regions, the SROCC report focused on herders instead. Following this approach, we designed our Google Scholar search accordingly and carried it out between May and July 2020. Our search method approximates that used by Petzold et al. (2020) and others, adapted to the limitations imposed by our lack of external resources.

Our criteria for inclusion at this stage were strict. We only considered papers in peer-reviewed scientific journals, so that books and book chapters were not included. This choice was made because the degree to which books are peer reviewed is often unclear. We also included only papers published since 2010. And we included only papers that report original results about Indigenous knowledge and the cryosphere. Many of the papers found in our literature search, including some of our own, are commentaries on topics related to Indigenous knowledge, such as the degree to which Indigenous and scientific knowledge overlap or connect. While such papers are important for other reasons, they are not directly relevant for the purposes of the present paper and were thus excluded. Our criteria also meant that some of the publications included in the list we sent to the SROCC authors were removed at this stage. In one way, we were surprised at the number of papers we found, which show a robust academic interest in Indigenous knowledge and a strong commitment by many scholars to publishing the results of their studies in peer-reviewed journals. In another way, we were surprised at the number of papers that we were aware of that did not meet our criteria. Their absence may also reflect limitations of our search process.

For Russia we focused on the words: Russia + climate change + traditional knowledge. We found 4,240 results, but most of these papers did not focus on cryosphere issues affecting Indigenous Peoples. The search was narrowed to focus on: Russia + traditional knowledge + reindeer herding + snow, which produced 496 results. The next search was Russia + traditional knowledge + reindeer herding + permafrost, which returned 279 results. An additional search was conducted using the term Indigenous knowledge as follows: Russia + Indigenous knowledge + reindeer herding + snow 301 results. The last search was Russia + Indigenous knowledge + reindeer herding + permafrost, which yielded 169 results. A search was conducted for glaciers using the following words: Russia + traditional knowledge + reindeer herding + glaciers, which returned 160 results. An additional search was conducted using the term Indigenous knowledge as follows: Russia + Indigenous knowledge + reindeer herding + glaciers, which gave 92 results. After reviewing the first two pages of results for the latter two searches, we determined that none of the papers specifically addressed glaciers, but instead focused in general on the Sami or on northern communities.

From these results, we identified 16 papers that met the criteria of addressing cryosphere-related changes in the Arctic and including Indigenous knowledge gathered by field work or interviews, not simply referring to it from other papers or documents or as a matter of policy rather than original research. Most of the results referred to papers that were government reports and assessment, theses, or other documents that were not peered reviewed, or publications in political papers or scientific publications that directly include traditional, Indigenous or local knowledge, or they were not in English. In other cases, the results referred to papers in which the key words appeared only in the references cited.

The same approach was followed for Fennoscandia as follows: Fennoscandia + traditional knowledge + reindeer herding + snow + ice, which produced 45 results. The next search included the following words: Fennoscandia + traditional knowledge + reindeer herding + permafrost, which returned 74 results. Fennoscandia + Indigenous knowledge + reindeer herding + snow + ice brought us 79 results. A search focusing on Fennoscandia + Indigenous knowledge + reindeer herding + permafrost, resulted in 53 results. The next search included the words: Fennoscandia + traditional knowledge + reindeer herding + glaciers, which returned 35 results, and the search for the words: Fennoscandia + Indigenous knowledge + reindeer herding + glaciers brought us 44 results. Using the same selection approach as was done for the Russian results, we found no additional papers that met our criteria.

For Greenland, where the cryosphere changes affect fishing and hunting for walrus and narwhal, whales, seals, and polar bears, our search included similar terms: Greenland + Inuit + traditional knowledge + snow + sea ice. The search produced in 1,090 results. An additional search, Greenland + Inuit + Indigenous knowledge + snow + sea ice, gave us 442 results. Of these, seven met our criteria for inclusion. The terms: Greenland + Inuit + Traditional Knowledge + glaciers, returned 761 results. Looking at the first three pages of results, two papers were already on our list and an additional one was identified that met our criteria. Searching for: Greenland + Inuit + Indigenous knowledge + glaciers returned 449 results. These results were similar to the previous search. No additional publications were identified when looking at the first three pages of results.

For Alaska the search terms Alaska + Inuit + traditional knowledge + snow + sea ice, producing 965 results. A second search, Alaska + Inuit + traditional knowledge + permafrost, gave us 761 results. An additional search was conducted, Alaska + Inuit + Indigenous knowledge + snow + sea ice, yielding 976 results. The search for Alaska + Inuit + Indigenous knowledge + permafrost returned 526 results. After reviewing these papers, we selected 19 that met our criteria for inclusion in the review. A third search, Alaska + Inuit + traditional knowledge + glaciers, gave us 1,300 results. An additional search was conducted, Alaska + Inuit + Indigenous knowledge + glaciers, yielding 887 results. No additional results were identified in the first three pages on either search. The next search removed the word Inuit and just focusing on: Alaska + glaciers + traditional knowledge. This search showed 1,370 results. No additional papers meeting our criteria were identified in the first three pages of google search.

For Canada, the search terms were Canada + Inuit + traditional knowledge + snow + sea ice, giving 1,160 results. The

TABLE 1 | Number of papers in the sample, by country and by region of Canada, showing coverage of cryosphere (equal to the total number of papers), additional topics, and responses to change.

Country/Region	Cryosphere (total papers)	Additional topics	Responses to change
Alaska	20	19	20
Canada	27	24	27
Fennoscandia	8	8	8
Greenland	8	8	8
Russia	13	13	12
Canada-wide	3	2	3
Nunatsiavut	7	7	7
Nunavik	1	8	9
Nunavut	9	1	1
Western Arctic	7	6	7

search for Canada + Inuit + traditional knowledge + permafrost produced 1,080 results. An additional search, Canada + Inuit + Indigenous knowledge + snow + sea ice, returned 767 results, and the search Canada + Inuit + Indigenous knowledge + permafrost found 692 results. From these results we identified 24 papers that met our criteria. A search followed for: Canada + Inuit + traditional knowledge + glacier. This search produced 554 results. An additional search was conducted for Canada + Inuit + Indigenous knowledge + glaciers, yielding 383 results. Once again, no additional papers meeting our criteria were identified in the first three pages of the results.

The total number of papers found was 76, which included those already compiled for the SROCC that met our new and more restrictive criteria. The final total also includes a few papers added through additional recommendations from colleagues. The references for all the papers in our sample is provided in the **Supplemental Material** of this paper, along with tables showing which papers were placed into which categories for our analyses of cryosphere coverage, additional topics, and responses to change. The **Supplemental Material** also provide a breakdown of which papers discuss which features of each component of the cryosphere, with graphs showing regional patterns, as described in section Patterns in the Coverage of Cryosphere below.

A summary of the number of papers and the topics covered, by country and by region of Canada (which had the highest total), is presented in **Table 1**.

Analysis and Limitations

All papers addressed at least one component of the cryosphere. From the descriptions in the papers, we identified several features of each component of the cryosphere and noted which papers addressed which features. The number of features identified ranged from two (glaciers) to ten (snow), depending on the amount of detail collectively provided by the papers that addressed each cryosphere component.

We next did a word search of the papers in our sample, using terms such as culture/cultural, society/social, economics, health, mental health, and policy/governance. Some of these terms had emerged in our previous work on the Adaptation Actions for a Changing Arctic report (Huntington and Eerkes-Medrano, 2017), for which we compiled a variety of stakeholder perspectives about changes in the Bering-Chukchi-Beaufort region of the Arctic. In addition to describing changes in the physical and biological environment, contributors to that report from Arctic communities discussed political, social, and economic factors affecting their lives. Looking more closely at those contributions and at the papers in our sample, we added culture, health, and mental health to the list of categories. Starting with the word search, we examined the context of the terms where they were found in each paper to confirm that we were assigning the papers to the right categories (e.g., that the use of "health" was not a passing comment or a reference to wildlife, but in fact a discussion of the topic with regard to humans). We define the topical categories used in section Patterns in the Coverage of Additional Topics as follows:

- *Social*: Issues relating to Indigenous people adjusting their hunting and gathering practices in response to change, including new techniques in the practice of their activities or the need to learn new ways to harvest, implying a social change in their activities.
- *Economic*: Effects on Indigenous activities related to financial costs and resources, such as having to use more fuel or to purchase larger boats to deal with increasing wave action or a change in reindeer practices that favor financial rewards from meat production rather than the aims of traditional husbandry.
- *Culture*: Changes that affect the ability to pass on Indigenous knowledge, such as use of new technology, new or adjusted hunting practices, loss of language, and others.
- *Policy*: Government regulations, policies, or practices that affect the practice of traditional activities.
- *Health*: Human health issues such a disease, heat-related concerns, respiratory issues due to air pollution or other physiological effects related to climate, diet, activities, and more.
- *Mental Health*: Issues resulting from the "sense of place" or "sense of worth" that is attached to the activities of Indigenous Peoples, as well as issues related to lack of housing or changes in culture or income.

In addition, we looked for discussions of responses, to understand how people in Arctic communities are reacting to the changes they see. This idea came again from many observations and comments by Arctic residents to us in our work over the years that adaptation is necessary and not a matter of choice. We also considered the ways in which the topic was discussed, for example how responses were described. We identified six categories of response, which are used in section Responding to Change:

- *Use IK*: Relying on Indigenous knowledge both for specific factual information and for more general attitudes about a healthy mindset for being productive. In some cases, papers emphasized sustaining IK, and others called for restoration or revitalization of IK.

- *Acknowledge variability and flexibility*: The awareness that the environment has always been variable, and that Arctic Peoples have long cultivated the flexibility to deal with that variability.
- *Shifts in practices*: Ways that people are or could be adjusting their hunting, fishing, gathering, and traveling practices in response to environmental change. This may mean using new times or areas or techniques, or learning to harvest different species entirely.
- *Cooperation with outsiders*: Working with government agencies and others to develop effective responses to change. This is a recognition of the limits of autonomous response (e.g., Huntington et al., 2017) and also often a call for partnerships in solving problems rather than a request to have problems solved for Arctic communities.
- *Cooperation with scientists*: A specific form of cooperation with outsiders, involving two-way sharing of information to help develop more effective responses. Again, this often emphasizes equal partnerships rather than visiting scientists making the decisions about what to study and how, or providing unilateral advice.
- *Sharing innovations*: Learning from one another as individuals and communities in the Arctic to develop new ways of doing things. Here the emphasis is on learning from peers, including those who may already be using species or techniques that are newly relevant in other areas.

Our sample of 76 papers provides the basis for our analysis of topics and patterns. We recognize, however, that this sample has a number of biases and limitations. First, only peer-reviewed scientific journal articles are included, which means a great number of other sources are excluded. Our search was limited to English-language publications, which likely excludes a number of papers, especially from Russia and also by scholars working in other languages elsewhere. The use of other search engines instead of or in addition to Google Scholar may produce different results, too. We have no reason to think that the peer-reviewed publications are a random sample of the topics and areas where studies have been done, and even less reason to think that existing studies provide an accurate and comprehensive picture of the views of Arctic Indigenous Peoples. Our conclusions thus concern only what has been reported in a particular fashion, and are more than likely to omit a great deal more that has been written and said in other ways. We do, however, believe that our sample, though certainly not exhaustive, is reasonably representative of the peer-reviewed literature about Indigenous knowledge and the Arctic cryosphere.

Second, published papers appear late in the lifespan of a research project or after the project itself has been completed. Our sample thus does not include many recent or current research efforts, and thus is inevitably out of date with regard to current patterns in research topics and areas. We are not aware of any major changes in the direction of research involving Arctic Indigenous Peoples, but the field is constantly evolving.

Third, our sample started with the criterion that papers address some aspect of a changing Arctic cryosphere. This topic is only a small portion of the research being done in and with Arctic communities. Our finding that there is a dearth of studies examining the mental health aspects of a changing cryosphere does not necessarily mean there is a dearth of mental health studies in the Arctic. An assessment of the adequacy of such studies in general is beyond the scope of this paper. Our concern is solely with the degree to which studies involving the changing cryosphere have made connections to related aspects of that topic.

Fourth, the focus on Inuit in Greenland and North America and on herders in Eurasia limits the degree to which the results are comparable, especially in terms of covering the various components of the cryosphere. Inland communities in North America, including Dene in Canada's Northwest Territories, First Nations in Canada's Yukon, and Athabascans in Alaska, are poorly represented in our sample and in the SROCC chapter. These communities are also affected by changes in permafrost, fresh water and lake ice, snow cover, and glaciers, though typically not to the same degree as the coastal communities who rely on sea ice. Similarly, coastal communities in Chukotka, Russia, are not represented, nor are the Sea Sami of northern Norway. Further research is needed to determine how much documented material related to the cryosphere is available concerning these Peoples. With regard to additional topics and responses to change, we believe comparability is greater, but should still be treated with caution.

We note that these limitations stem primarily from the terms imposed on authors of IPCC reports, which rely, with few exceptions, on publications that have appeared in the peer-reviewed literature by a cutoff date set for each IPCC report. The SROCC chapter to which we contributed focused on the cryosphere changes and its impacts on particularly vulnerable human populations, limiting the range of papers that were appropriate for consideration. The limitations of our study, therefore, are likely to match closely those of any IPCC report or any reports using similar criteria for identifying acceptable evidence. A wider-ranging review of research involving Arctic Indigenous Peoples will no doubt identify a great many more papers in each of our categories, but we expect that the vast majority of those additional papers will lack an explicit link to the changing cryosphere.

RESULTS

Patterns in the Coverage of Cryosphere

The combined searches for recent papers concerning Indigenous and traditional knowledge of the Arctic cryosphere yielded a total of 76 that met the IPCC criteria for consideration (not including the SROCC cut-off date) and that substantively engaged with Indigenous and traditional knowledge. Since the search emphasized environmental terms, it is not surprising that all of the papers consider environmental issues. The various cryosphere components were not covered equally in all countries (considering Fennoscandia as a unit on par with the others; **Figure 2**). Not surprisingly, sea ice did not come up in Fennoscandia and only rarely in Russia, as our search there focused on herders, and there are few coastal communities with extensive use of sea ice. Observations concerning freshwater were not documented in Greenland. Permafrost was not mentioned in

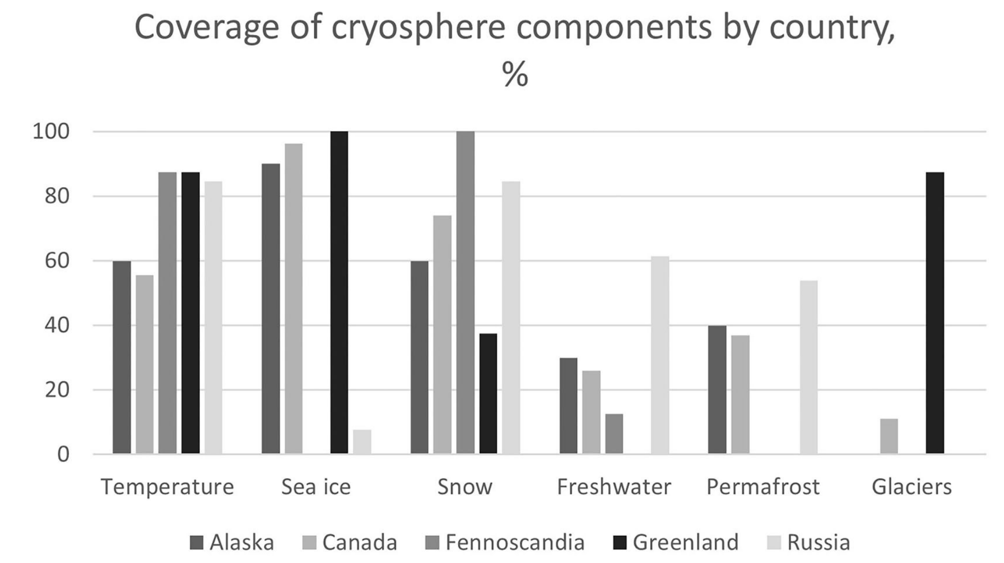

FIGURE 2 | Coverage of cryosphere components by country, % of total papers.

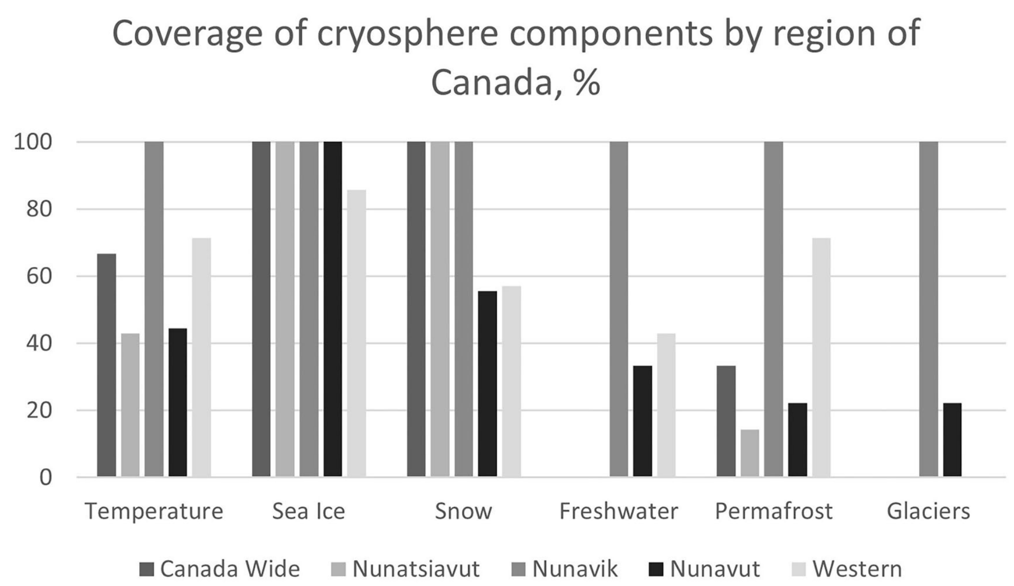

FIGURE 3 | Coverage of cryosphere components by region of Canada, % of total papers.

either Fennoscandia or Greenland. Glaciers were mentioned only in Canada and Greenland.

Given the size of the Canadian Arctic and the number of papers from Canada, we separated the Canadian papers by region into Nunatsiavut, Nunavik, Nunavut, and the Western Arctic (including the Inuvialuit Settlement Region and the northern Yukon). With the exception of one paper from the Western Arctic, all the papers were focused on Inuit and Inuvialuit. The exception examined the Vuntut Gwich'in from Old Crow, Yukon. As with the circumpolar overview, the treatment of cryosphere components (**Figure 3**) varies by region. Not surprisingly, sea ice and snow were covered in the majority of papers.

We also examined the degree to which different features of each cryosphere component were discussed or described in the papers in our sample. For sea ice, thickness, extent, freezeup, breakup, and shorefast ice received the most attention, among countries and regions of Canada. For snow, depth/thickness, less snow, and different timing were the most noted features across countries. For Canada, the order of coverage among those three features was reversed. Fennoscandia stands out for a high level of interest in compaction, different snow characteristics, ice-on-snow events, and late snow, which received little or no attention elsewhere. For freshwater, river break-up, shallow rivers, and thin ice were the most widely noted features. Among

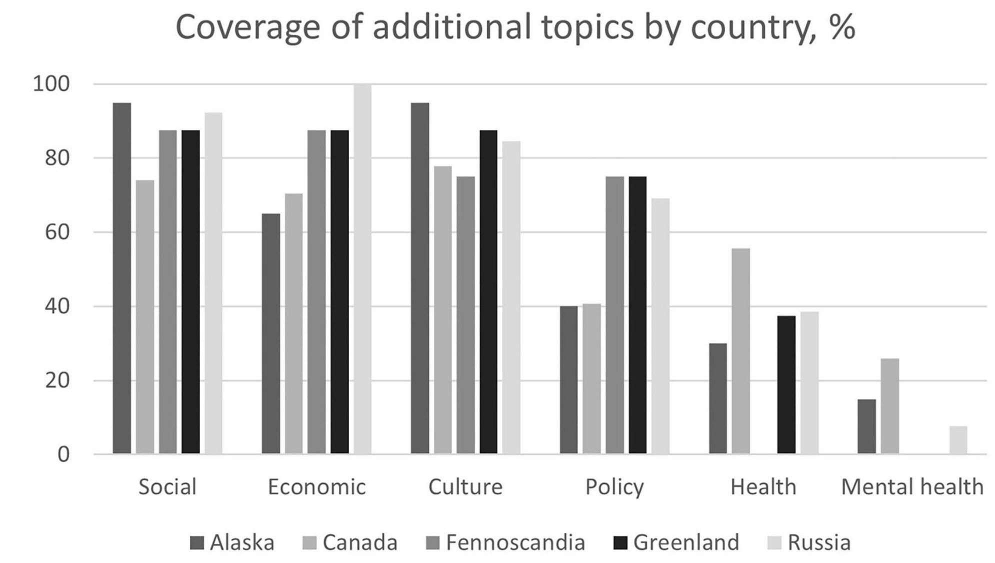

FIGURE 4 | Coverage of additional topics by country, % of total papers.

the three countries for which permafrost observations were reported, thawing and warming were most prevalent, followed by stability and thermokarst. The latter two were not mentioned in Canada, and thermokarst was only mentioned in Russia. Glaciers received the most coverage in Greenland, where all of the papers addressed glacial retreat. Three papers mentioned glaciers in Canada, two of which discussed effects from glacial melt, and one of which discussed glacier retreat. The details of the coverage of cryosphere features are provided in the **Supplemental Material** accompanying this paper.

Patterns in the Coverage of Additional Topics

What is interesting for the purpose of this paper is the degree to which other topics are included (**Figure 4**). Social issues were mentioned most frequently, closely followed by cultural issues and economic issues. Policy, health, and mental health were also included, though to a lesser degree. At the national level, different patterns emerge. All the Russian papers mentioned social issues, as did all but one of the papers concerning Alaska and all but one concerning Fennoscandia. A smaller majority of the papers from Canada and Greenland also mentioned social issues. The pattern was broadly similar for cultural issues. For economic issues, the proportion of papers in Alaska dropped, but remained similar elsewhere. Policy, on the other hand, was mentioned in fewer than half of the Alaska and Canadian papers, but in most of the papers from Russia and Fennoscandia, and all of the Greenland ones. Health received the most attention in Canada, followed by Russia, whereas mental health was only addressed in papers from Canada and from Alaska, with the former having twice as many as the latter.

In Canada (**Figure 5**), papers from the Western Arctic were less likely to make connections between the cryosphere and other topic areas. Mental health was addressed primarily in Nunatsiavut, with an additional mention in one paper about the Canadian Arctic generally. Policy and health were addressed more widely, but still at a lower overall level than social, economic, and cultural matters.

Responding to Change

In addition to discussing the connections between a changing cryosphere and other topics, most of the papers also described one or more response actions being taken or being suggested by the people in the study areas.

There appears to be a widespread recognition of all six response categories (**Figure 6**), though considerable variation by country. Papers from Alaska, for example, have infrequently described the acknowledgment of variability or the sharing of innovations, whereas papers about Russia are most likely to identify acknowledging variability as a response. Overall, and acknowledging the small sample size, papers about Greenland are most likely to refer to a wide range of responses. These differences may be an artifact of the relatively small sample sizes, or a reflection of our different search foci in the Inuit region and in Eurasia, or of differences in the societal and geographical contexts of the different regions.

Within Canada, there is again regional variation (**Figure 7**). The use of Indigenous knowledge is widespread, as is cooperation with scientists. Papers from Nunatsiavut and the Western Arctic were less likely to report responses to change than papers from Nunavik and Nunavut, though again the small sample size may be a factor.

DISCUSSION

Climate change does not occur in isolation. A changing cryosphere affects Arctic Indigenous Peoples across many intersections of social and cultural change, economic challenges,

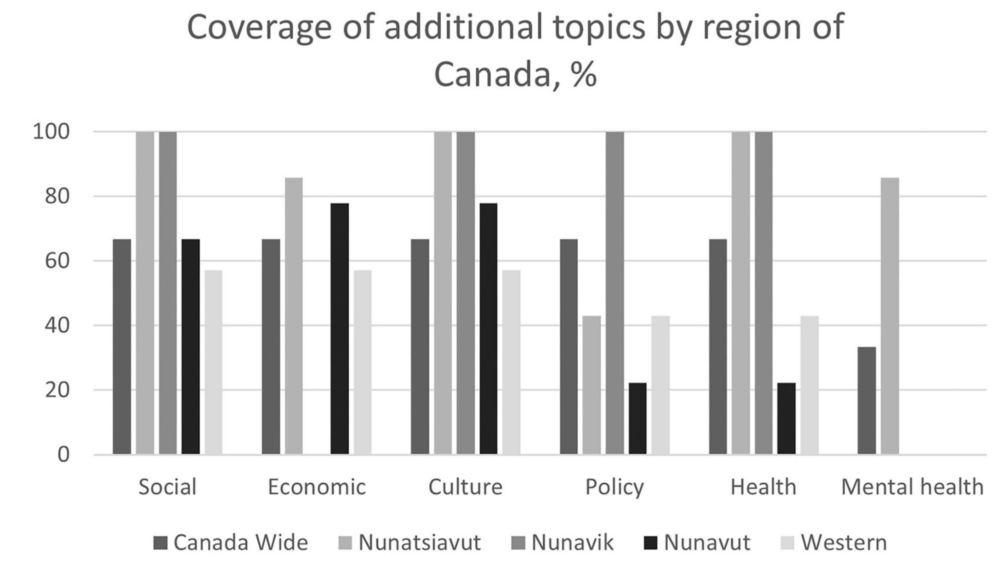

FIGURE 5 | Coverage of additional topics by region of Canada, % of total papers.

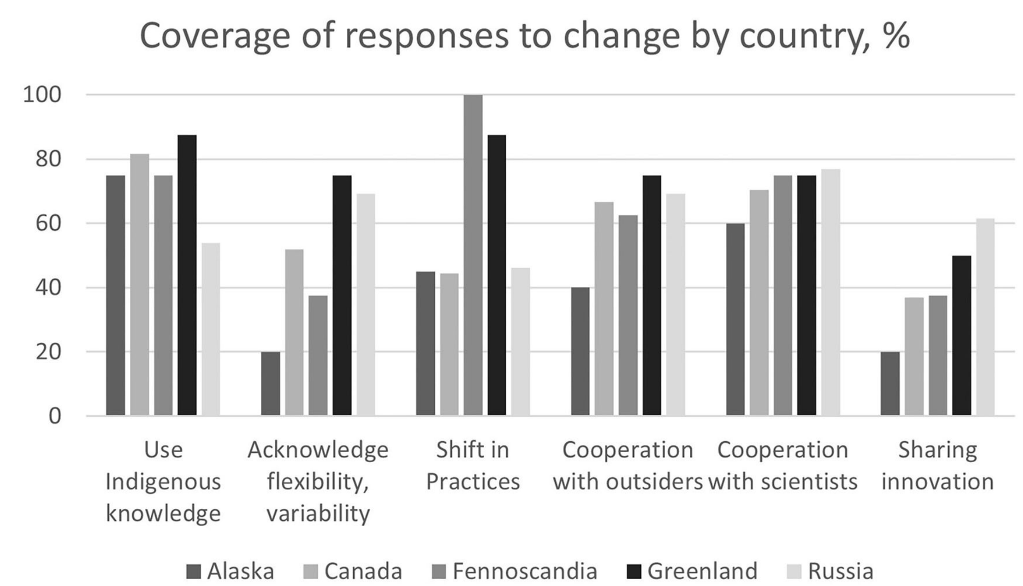

FIGURE 6 | Coverage of responses to change by country, % of total papers.

limited access to services such as health care, and more. Climate change is also not the whole story. As far-reaching as the effects of thawing permafrost, melting sea ice, and reduced snow cover are, human responses still have great scope to determine the ultimate outcomes of these changes. We now discuss the themes of cryosphere, other topics, responses to change, and what we believe to be the "untold stories" of a changing cryosphere.

Cryosphere Coverage

As seen in **Figures 2**, **3**, the different elements of the Arctic cryosphere and the inclusion of Indigenous knowledge in identifying impacts on Indigenous people are all addressed, though to varying degrees in different countries and across the regions of Canada. When we carried out our literature search for the SROCC, we were pleasantly surprised at the range of papers from all around the Arctic that document and discuss Indigenous knowledge about the changing cryosphere, even with the limiting criteria we used. Perhaps this is a reflection of low expectations on our part, but nonetheless our findings for the SROCC and the additional papers presented here offer a strong counterargument to claims that Indigenous knowledge is missing or sparse in the peer-reviewed literature (e.g., Arsenault et al., 2019; Petzold et al.,

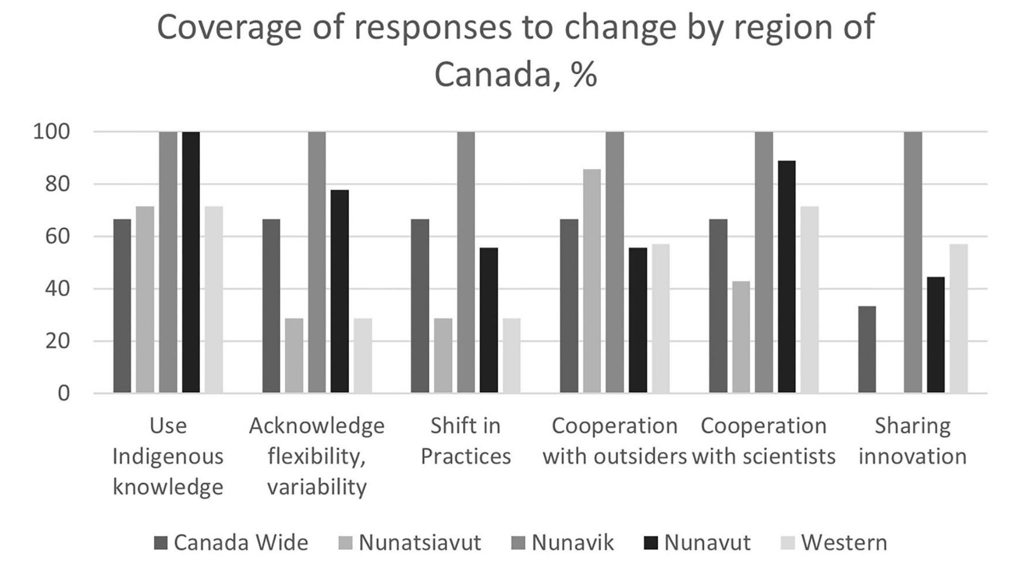

FIGURE 7 | Coverage of responses to change by region of Canada, % of total papers.

2020). Even 76 papers, however, fall far short of the available peer-reviewed literature from the natural sciences. Thus, it appears fair to state that Indigenous knowledge remains underrepresented in scientific publishing, greatly limiting its availability to the work of the IPCC and other bodies that rely primarily or exclusively on peer-reviewed scholarly publications.

We should also not be surprised by some of the patterns in coverage from country to country. The Sami of Fennoscandia have little interaction with sea ice, so it is not surprising that although the SROCC has an in-depth analysis of sea ice and impacts on aquatic ecosystems, fishers, tourism, and other topics, no studies from that area address changes in sea ice. Most Indigenous Peoples in the Russian Federation are not maritime, either, so again it is not surprising that there are few peer-reviewed papers looking at knowledge of and experiences with sea ice in Russia. Similarly, ice-free ground in Greenland is typically rocky, so there are likely to be fewer concerns about permafrost than would be in regions such as western Canada and Alaska, where thawing ground is likely to turn to mud or water threatening some coastal and riverbank communities especially in Alaska (e.g., Melvin et al., 2017; Meredith and Sommerkorn, 2019), or in Russia where the thermokarst lakes are also disappearing, transforming the grassland landscape required for herding activities (Istomin and Habeck, 2016; Crate et al., 2017). More surprising perhaps is the comparative dearth of studies about snow in Alaska that met our criteria, and about freshwater anywhere outside of Russia, though the latter may be a reflection of the focus on herders and the lack of corresponding inland information from North America. Both suggest opportunities to fill geographical gaps in the documented record of the changing Arctic cryosphere from Indigenous Peoples' perspectives or to expand the scope of the study to include all communities throughout the Arctic. Of the cryosphere components we examined, permafrost has received the least attention, again suggesting a gap to be filled. Even in Russia, where the prevalence of permafrost-related studies is highest, only 7 of 13 papers address it.

A finer-scale analysis could also be done of cryosphere coverage, both in terms of geography (e.g., how much of Russia do the seven permafrost papers actually cover?) and depth of treatment (e.g., what aspects of snow are actually addressed and in which locations?). We cannot anticipate all of the detailed questions that cryosphere researchers might ask, and thus cannot provide here all of the details and nuances that might be desired. Nonetheless, we can at least point out that studies documenting Indigenous knowledge and use of sea ice and snow typically consider many specific aspects of those topics. For example, studies of sea ice and snow terminology illustrate the aspects of each that matter to people, depending on the activities they are engaged in. Marine mammal hunters are concerned with safety (cracks, thin ice, potential for shorefast ice to break off), with animal habitat (where breathing holes or lairs may be, the types of ice floes where marine mammals are likely to be found), with platforms for butchering animals (thickness of shorefast ice for large whales, stability of ice floes for walrus and bearded seals), and other aspects of sea ice. Reindeer herders are concerned with travel (snow depth and crust strength), the animals' ability to reach food under the snow (rain-on-snow events or icing at the bottom of the snowpack), the way a breakable crust may damage the legs of the animals, and other aspects of snow. Of course, Indigenous knowledge is not strictly utilitarian, either, so a great deal more will be known to those who spend their lives on snow and ice. We also note that those documenting Indigenous knowledge cannot anticipate what other researchers will be interested in, so a great deal remains undocumented even in places where studies have already been conducted.

While there is a good analysis of permafrost issues from the physical sciences perspective in the SROCC, connections between permafrost and Indigenous Peoples are largely absent. Such an analysis is beyond the scope of this paper, but could be a fruitful line of inquiry in support of the next major assessment of the changing Arctic cryosphere and impacts on Indigenous Peoples, as well as to allocate research funding to fill specific gaps in addition to addressing major geographical ones. We note also the difficulty of determining whether the lack of coverage reflects a lack of knowledge or just a lack of scholarly inquiry and writing. For example, permafrost is present in Fennoscandia, but in the absence of further information, we cannot tell whether thawing permafrost simply does not affect Sami to any degree, or whether no researchers have asked and gone on to publish the answers. Even if the answers are negative, the questions are worth asking to make sure a substantial body of knowledge is not simply being ignored.

Additional Topics

The connections between the cryosphere and other aspects of life in the Arctic make it all but inevitable that discussions about a changing cryosphere will entrain other topics as well. As can be seen in **Figures 4**, **5**, a wide range of such topics is addressed in the papers in our collection. Social, cultural, and economic issues are widely discussed, as these are broad categories (as we use the terms) encompassing a wide range of outcomes and influences. Recognizing that some of the differences could be an artifact of our classification, it is nonetheless interesting to note that in Canada only three-quarters of the papers mentioned economics, with the Western Arctic region least likely to do so. These differences may reflect different systems of land access and management across the four regions of the Canadian Arctic, resulting in different priorities and perspectives from region to region. The result could also be an artifact of the types of studies done in that region, rather than an actual lack of economic aspects of a changing cryosphere.

Among the other three topics, policy received the least coverage in Alaska, while health generally lagged other topics, and mental health had the least coverage in all regions. As with economics in the Western Arctic of Canada, we cannot tell if the relative lack of attention to policy in Alaska is a result of where researchers chose to focus their efforts, or a systemic disconnect between policy and cryosphere. The latter seems unlikely, given the attention to coastal erosion and sea ice loss among other topics in Alaska (e.g., Marino, 2015; Melvin et al., 2017), but more work is required to understand what may be missing. Mental health is perhaps the clearest-cut case of researcher influence. The topic is covered to the greatest extent in Canada, and even there in only one quarter of the papers considered. Of the six papers addressing mental health in Canada, five are from Nunatsiavut, and one researcher (Ashlee Cunsolo) was a co-author on all six. We consider it highly unlikely that the connections between the cryosphere and mental health are most pronounced in Nunatsiavut, and far more probable that the uneven distribution of mental health papers is a result of Cunsolo's leadership in studying this topic. In other regions and for other topics, we saw a greater range of researchers involved in studying Indigenous knowledge and the cryosphere. The participation of more researchers, however, does not necessarily mean that more topics are covered, as researchers may well continue along similar paths as their predecessors and colleagues.

A comprehensive understanding of the ramifications of a changing cryosphere for Arctic Indigenous communities requires more than chance attention to related topics. As is the case for the cryosphere components themselves, there are major geographical and other gaps in the treatment of additional topics. The most notable, if not the most pressing, gap is the lack of attention to mental health. Other topics, too, deserve a more thorough discussion across countries and regions to identify with greater confidence the patterns in how a changing cryosphere affects community and individual well-being.

We also recognize that our topics are broad and, to some extent, overlapping. Further analysis is needed to assess the degree to which the various topics are addressed fully. For example, "economics" spans income, employment, comparative costs, taxation structures, subsidies, development, and many other aspects of access to money and other resources. To say that many papers discuss economics does not in any way mean that economics has been fully covered in any country or region. Further work would be needed to examine any of these topics in detail. Our findings simply show that Arctic Indigenous Peoples' views on issues other than climate change are relevant and are starting to be reflected in the scientific literature.

As noted in section Analysis and Limitations, our sample of 76 papers does not include a great many more papers addressing various aspects of life in Arctic Indigenous communities. An additional area for future work is to cast a wider net than cryosphere-related papers, to consider for example the degree to which mental health or economics or other topics have been addressed overall around the Arctic. Additional scholarship would then be needed to connect a potential wider body of work on any of these topics with the cryosphere-related work examined here. Doing so could add a great deal of depth to our understanding of the context in which cryosphere change affects Arctic Indigenous Peoples, and thus form the basis for a much more thorough examination of such issues in future assessments of the effects of a changing cryosphere.

Responding to Change

Concerning the various responses to change, we have to keep in mind that Arctic Indigenous Peoples have a history of adaptation from long before the current environmental changes they are facing (Krupnik, 1993). Indigenous Peoples have inhabited these regions for thousands of years (Coates and Broderstad, 2020). For example, the Bering-Chukchi-Beaufort region has been inhabited for more than 10,000 years (e.g., Anderson, 1988). Commercial exploitation of the region's resources began in 1848 with the hunt of bowhead whales by Yankee whalers (Haycox, 2020). Subsequent activities included walrus hunting and fur trapping (Bockstoce, 2010), the development of the Northern Sea Route (AMSA, 2009), oil and gas exploration (AMAP, 2010), construction of national defense systems such as the Distant Early Warning (DEW) Line across northern North America (Jenness, 1962), mining, tourism, and commercial fishing (Arnold et al., 2011). These economic developments have been accompanied by far-reaching social and political changes, from the exercise of national jurisdiction throughout the area by Canada, Russia, and the United States; to the settlement of Indigenous land claims in Alaska and Canada and the creation of local governments in northern Alaska and self-rule government in Greenland (Nuttall, 2008; Zellen, 2020); from the advent of modern technology and communications to the ongoing loss of Indigenous languages (Barry et al., 2013).

In this context, recent environmental change is seen by many Indigenous persons as simply one of many forms of change they are experiencing. For example, a study by Johnson et al. (2016) outlines that to date Inuit have been able to adapt to thinning ice the best they can, but their concern is the potential change from increased shipping activities and northern development. In Russia, an additional factor affecting them is the lack of mobility to their ancestral lands, as well as changes in subsidies and government support (Nakada, 2015). In the Barents region, the Sami have always adapted to weather conditions but now they are affected by government regulations and the loss of land for agriculture, forestry, mining industry, construction of dams for power generation, tourism, and new market economy (Ksenofontov et al., 2017; Kirchner, 2020). In addition to the effects of cryosphere change on traditional activities, further attention may be warranted on all aspect of life in Indigenous and other Arctic communities. For example, thawing permafrost can damage infrastructure such as roads and buildings, reducing quality of life, and even threatening health and safety. Studies of Indigenous knowledge tend to focus on traditional activities, but these are only part of today's lives and livelihoods in the Arctic. More work is needed to make an explicit connection between Indigenous knowledge, a changing cryosphere, and the full range of today's activities and concerns in the Arctic.

In our review, we use the term of "responses to change" as a non-judgmental alternative to "adaptations," and present it as an additional category to be explored. Much has been done on the topic of responses to climate change, often using approaches such as adaptation (Berkes and Jolly, 2002; Armitage et al., 2011; Ford et al., 2014, 2015; AMAP, 2017a,b,c), resilience (Crane, 2010; Forbes, 2013; Ford et al., 2020), and vulnerability (Ford and Smit, 2004; Ford et al., 2008; Keskitalo, 2016). The types of response we have identified are intended mainly to illustrate the range of options being pursued (**Figures 6**, **7**), noting many other attempts to categorize responses and response mechanisms (e.g., Thornton and Manasfi, 2010; Walker and Salt, 2012). What we find most noteworthy is the divergence between a popular narrative of gloom and doom in the Arctic (e.g., Van Tuyn, 2013; Herrmann, 2018) and the descriptions of many robust mechanisms being pursued or suggested by Arctic Indigenous Peoples today.

The research reported in the papers in our collection did not necessarily set out to document responses to change, just as the studies did not necessarily set out to address topics beyond the cryosphere, either. Nonetheless, many papers have much to say in both areas. Regarding responses to a changing cryosphere, some patterns stand out. Overall, the use of Indigenous knowledge is the most common response, closely followed by cooperation with scientists. The latter is perhaps not surprising in papers written by scientists, though we recognize that information from scientific work is valued in many Indigenous communities. Cooperation with outsiders is most commonly reported in Greenland and Russia, as is, to a lesser degree, acknowledgment of variability, and flexibility. We are reluctant to read too much into these differences, recognizing the potential for artifact in a small sample, but further work might explore the role of societal context as well as researcher interests.

Untold Stories and Their Relevance to Future IPCC Reports

In addition to what we have examined so far, we noted another pattern in many of the papers in our sample. Beyond the general results presented, we came across many personal stories of adaptation reflecting that, for Indigenous Peoples, adaptation to environmental change is a constant in their lives and they will continue to adapt. These stories stand in contrast to many reports of doom and gloom from the Arctic, in both the popular media (e.g., Herrmann, 2018) as well as in scholarly publications (e.g., Van Tuyn, 2013; Moon et al., 2019). While these views may not quite be "untold," they appear not to be widely recognized or acknowledged. Future assessments of a changing Arctic could benefit from including personal stories by Indigenous individuals that convey what people are experiencing and what those experiences mean for them and their communities. We do not suggest that such stories should be given greater weight than the results of broad surveys that are likely to be more representative of large groups, but rather that stories can illuminate details and nuances that may be essential to the accurate interpretation of broad patterns. Furthermore, the tone of such stories can provide valuable insight into attitudes toward change and response, attitudes that may themselves make the difference between lasting hardship and successful adaptation.

To illustrate our point, we present some of the stories from different regions of the Arctic.

In Greenland, Indigenous people have seen over their lifetimes varying sea ice conditions including periods of heavy ice but also light or even no ice. For example, near Aasiaat in the 1940s, the sea ice was thin and unstable and people could not travel by dog team to villages in the south (Niaqornaarsuk and Iginniarfik). They had to use their boats to get to southern locations, and in the winter of 1946–1947, the area had no sea ice (Holm, 2010). In contrast, years later, there were extreme low temperatures and the sea ice formed fully. Erik Røde Frederiksen, a sheep farmer interviewed in 2006, mentioned that their forefathers used to say that if it is calm now, it will be windy in the future, "eqiterpaageeq anerlertarnissaminut silaannaap qatsingarujussuarlini," "the weather is collecting the future winds by being calm," reflecting a cycle of weather and climate (Holm, 2010).

Sami herders in the Barents region talk about how they have adapted to cyclic variations in the natural environment. These herders consider that they are handling climatic variations better than other cultures because herding knowledge requires effective adaptation to rapid and unexpected changes and variable conditions (Reinert et al., 2008). A herder in the Kaldoaivi cooperative in the Barents region said: "I'm not at all afraid of climate change; for reindeer, it signals a good direction, until now" (Vuojala-Magga et al., 2011).

In Alaska (Huntington and Eerkes-Medrano, 2017), George Noongwook from Savoonga pointed out that he has been more concerned about the quota imposed on bowhead whalers by the International Whaling Commission (IWC) than about climate change. Other contributors to the same report were concerned about the impacts that oil and gas development in the Chukchi and Beaufort Seas may have on the whales and food sources due to pollution, noise, and other disturbances. In Shaktoolik, Eugene Asicksik mentioned that government regulations also pose more of a problem to the fishing industry than does climate change.

Similar views were identified in a study by the Inuit Circumpolar Council in 2014 (Johnson et al., 2016). Hunters from Alaska, Canada, Greenland, and Chukotka (Russia) were interviewed and their key messages reflect that Inuit have been adapting throughout their history to extreme fluctuations and conditions and are confident that they will succeed in adapting to changes in climate and sea ice because they are adaptable and strong. The uncertainties they feel are about future impacts of increased development and shipping activities.

In the case of Russia, political issues such as the treatment of Indigenous Peoples, policies concerning mineral and petroleum development, and the availability of government support may be more immediate concerns than climate change. Reindeer herders in the Oymyakon District report that the reindeer numbers have been increasing, as of the time of the study, but recent air temperatures have not been increasing, and so their concerns have more to do with government policies than with the climate (Nakada, 2015).

For many Indigenous Peoples, a key to adapting to changing conditions has been the ability to move freely. For some, this has now changed. For the Sami in Norway, Sweden, Finland, and Russia, government regulations and modern infrastructure limit the reindeer herders' ability to access ancestral and private lands for pasture or to alter their patterns in response to change (Vuojala-Magga et al., 2011). In Greenland, today's modern settlements have many conveniences but are not mobile, in contrast to seasonal camps used previously, and so Inuit have far less ability to move their camps to new hunting grounds according to their needs and the changing seasons. This lack of mobility makes them feel that changes are now more visible and affect them more profoundly than in the past (Holm, 2010).

The format and rules governing reports such as the SROCC do not leave much room for personal reflections. First, contributions are expected to be concise, and many quotes and stories use more words than may be available. Second, the expectation of defensible objectivity makes it difficult to include individual voices, even to illustrate more general points. Some reports have included short quotes as illustrative material (e.g., Meltofte, 2013), but we find this practice superficial rather than satisfactory. Furthermore, attempting to fit Indigenous knowledge and voices into a predetermined structure and topic is better than ignoring those voices, but still requires taking information from one context and using it in another. Providing space for Indigenous authors to present their own views in their own voices is one alternative, which has recently been used in the annual Arctic Report Cards issued by the National Oceanic and Atmospheric Administration in the U.S. (Slats et al., 2019).

The funding sources and motivations for climate change research and media coverage are also a factor in shaping the stories that are told. Efforts to document and report the magnitude of climate change have, understandably, tended to emphasize the scope of change and the risks that human communities and societies are facing. We do not wish in any way to detract from the excellent work that has been done or the seriousness of the topic. Nonetheless, as Arnold (2018) points out, information that does not fit the prevailing narrative often receives less attention. For example, the record sea ice minimum of 2007 received considerable media attention, but did not create a major impression on residents of the north coast of Alaska, closest to the newly open water, a fact that received far less coverage (Christensen et al., 2013). Part of the story of Arctic change is being told, but part remains obscured or missing. Further work is needed to remedy this imbalance.

CONCLUSIONS

Global attention to the changing Arctic cryosphere has helped spur interest in the knowledge and experiences of Arctic Indigenous Peoples on this topic. Through a combination of researcher interest and Indigenous self-expression, the 76 studies we found that examine a changing cryosphere from the Indigenous perspective also present connections to other aspects of community and individual life, describe a range of responses to change, and include some "untold stories." We are grateful to the authors of all of these papers for taking a wide look at their topics and providing a fuller view of life in a changing Arctic. It is particularly important that Indigenous voices are heard and that Indigenous ideas are part of the discussion about Arctic change and its meaning.

That said, we also recognize that scientists and journalists often focus on the evidence that matches their ideas (Arnold, 2018), rather than questioning their assumptions and the prevailing narrative. When it comes to the Arctic cryosphere, the prevailing narrative is often one of unstoppable, catastrophic change (e.g., Moon et al., 2019; Huntington et al., 2020). While a changing cryosphere does indeed affect people's lives in many ways large and small, that is far from the whole story. The papers in our collection collectively suggest a different narrative, one emphasizing connection and response, rather than inevitability and loss.

Out of thousands of scientific papers concerning the Arctic cryosphere, the ones that we found are both welcome and inadequate. Welcome, because they offer insight that is otherwise lacking, and a chance to see the Arctic through Indigenous eyes, even if indirectly. Inadequate, because there is so much more to be said. In the section Discussion above, we identified major gaps in the coverage of the cryosphere, additional topics, and responses to change. The gaps we note are mainly geographic, in that coverage is uneven around the Arctic, suggesting that much remains to be learned. We also note that further work should look more closely, to identify gaps at a finer scale, within topical categories and at sub-national levels, as we have begun to do here for Canada. Additional work is needed to examine the papers in our sample, and other sources as available and appropriate, in more detail concerning specific topics such as the nuances of cryosphere characteristics and change, or the many ways in which a changing cryosphere affects all aspects of today's life in Arctic Indigenous communities. We have merely indicated the range of topics covered in the papers in our sample, and further studies could go into much greater depth on any one of these points.

Our review has shown that climate change is not always the primary threat to Indigenous well-being in the Arctic, one spelling doom for Arctic Indigenous peoples. Their own narratives offer a contrasting story of response and resilience. We suggest that the IPCC continues its efforts in partnership with Indigenous peoples to identify thematic and regional gaps in coverage, and that the required time and funding is allocated to fill such gaps. Such an effort may also require recruiting additional researchers with the required expertise and providing opportunities for inter-regional information sharing among Arctic Indigenous Peoples. As part of this effort, more work should be put into providing platforms for telling the untold stories of Arctic change and Indigenous Peoples, so that change can be understood in the context of culture and history as well as that of climate and geography.

At present, we have the ability to tell a partial story, based on what a number of individual researchers have taken upon themselves, in cooperation with Indigenous communities around the Arctic. A more systematic effort is needed to develop a more complete, comprehensive story built from a concerted effort to address core topics consistently. Doing so will help us move scholarly documentation of Indigenous knowledge and experiences from an important but uneven contribution to an essential foundation for understanding and addressing a changing cryosphere, on Indigenous terms and in service of Indigenous aspirations for their own futures. The SROCC is a step in that direction, and future work can and should continue along this path.

DATA AVAILABILITY STATEMENT

The original contributions generated for the study are included in the article/**Supplementary Material**, further inquiries can be directed to the corresponding author/s.

AUTHOR CONTRIBUTIONS

LE-M and HH conceived of and wrote the paper. LE-M carried out the literature search and textual analysis. Both authors accountable for the content of the work.

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SUPPLEMENTARY MATERIAL

The Supplementary Material for this article can be found online at: https://www.frontiersin.org/articles/10.3389/fclim.2021.675805/full#supplementary-material

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IPCC and the Deep Sea: A Case for Deeper Knowledge

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IPCC reporting culture and structure leads to a failure to highlight potential vulnerabilities and risk in areas where research is largely absent. Nowhere is this more obvious than in treatment of the deep ocean (waters below 200 m), where climate research is in its infancy, but human exploitation of resources is on the rise. Understanding climate-induced changes in deep- sea environments, ecosystems and their services, including carbon cycling and climate regulation, is fundamental to future ocean sustainability and to decisions about active climate remediation.

Keywords: deep ocean, climate change, biodiversity, SROCC, UN Decade for Ocean Science

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INTRODUCTION

The Intergovernmental Panel on Climate Change (IPCC) reporting system for both the assessment reports (AR) and special reports (SR) provide assessment of climate change, its consequences for ecosystems, the implications for humans and possible solutions. It emphasizes information and topics for which there is a solid scholarly basis for assessment, often interpreted as multiple studies conducted by multiple research groups so that statistical likelihood or confidence language can be applied. The reports are concise, given the huge breadth of topics covered, thus text gets whittled down to meet stringent page limits, and summaries for policy makers (SPM) focus on conclusions with medium or high confidence. Under the auspices of IPCC Working Group 2 (Impacts, Adaptation and Vulnerability), I participated as one of many lead authors in preparation of the Special Report on Oceans and Cryosphere in a Changing Climate (SROCC), as a contributing author for AR5 and as a review editor for AR6. What these reports are not able to do is highlight potential vulnerabilities and risk in areas where research is largely absent. Nowhere is this more obvious than in treatment of the deep ocean (waters below 200 m).

DELVING DEEPER

Despite the fact that the deep ocean covers nearly half the planet's surface (and most of the habitable volume), and absorbs massive amounts of heat and carbon dioxide, direct observations of the consequences of climate change for modern deep-sea ecosystems are sparse (Levin and Le Bris, 2015; Sweetman et al., 2017). Nevertheless, earth system model outputs indicate that deep waters are warming, becoming more acidic, and loosing oxygen and that this trend will continue in the future under all emission reduction scenarios (Bindoff et al., 2019; Kwiatkowski et al., 2020). The projected consequences include species and productivity redistributions, habitat compression, biodiversity loss, and changes in body size, food webs and connectivity that can influence commercial harvest, carbon sequestration and nutrient cycling (Sweetman et al., 2017; Brito-Morales et al., 2020), but few *in situ* observations exist that can confirm these. The IPCC AR 5 does not address the deep ocean in detail and AR 6, while hosting two deep-sea subsections, has limited text thus far. The deep sea contains diverse ecosystems with different climate vulnerabilities. This emerges from the burning ember diagram in SROCC Ch. 5 (Fig. 5.16 in Bindoff et al., 2019), which notably offers risk assessment for 5 deep-sea ecosystems. But confidence levels are low, and thus three of the five ecosystems are not reflected in the SROCC SPM. Ecosystem heterogeneity in the deep sea is driven by reliance on photo- vs. chemosynthesis, topographic variation (e.g., canyons, slopes, ridges and seamounts), oligo-vs eutrophic settings, varied water masses and interactions with the epipelagic and mesopelagic realms (Ramirez-Llodra et al., 2010). The deep ocean contains both living and non-living energy and mineral resources of growing interest (Ramirez-Llodra et al., 2011; Mengerink et al., 2014). Despite this, early external reviewers of the SROCC Chapter 5 viewed the deep as one habitat and questioned why the deep "ecosystem" should get more space in the report than single coastal systems such as estuaries, sandy beaches or salt marshes.

The limited information about climate consequences for the deep ocean carries over to AR6. This is evident for example in the fact that open-ocean, deep-sea systems do not feature in the discussion in the AR6 WGII Cross Chapter paper on biodiversity hotspots. The 60% of the ocean which is deep barely registers in this discussion, despite knowledge of exceedingly high biodiversity (Rex and Etter, 2010), high endemism in settings such as hydrothermal vents and seamounts, and recognized climate vulnerability (Levin and Le Bris, 2015; Sweetman et al., 2017; Bindoff et al., 2019; Brito-Morales et al., 2020). Apparently needed quantification of biodiversity is missing.

Observational programs have generated a reasonable understanding of temperature change in the deep sea and Earth System model projections have been extended to the deep sea floor (Sweetman et al., 2017; Kwiatkowski et al., 2020), so there are now predictions of future seafloor exposures to changing temperature, oxygen, pH and more, the time of emergence of these signals, and other derived metrics. These can be used to project changes in species distribution or connectivity (e.g., Levin et al., 2020; Morato et al., 2020). But most of the information about faunal response to climate change in the deep ocean comes largely from the study of paleo records, natural gradients, very limited laboratory experimentation (e.g., for cold water corals) and basic biological knowledge of how living systems respond to temperature, rather than from extensive *in situ* observations over time or from laboratory experiments. There are only a handful of long-term study sites in the deep ocean that are capable of documenting modern ecosystem response to anthropogenic climate change. Those in the Pacific (Station M) and the Atlantic (PAP), exhibit an unanticipated level of temporal variability in ecological attributes that is difficult to attribute to secular climate change (Smith et al., 2013; Hartman et al., 2021). A limited number of observatories now have multi-decadal data, but these monitor just a few types of ecosystems (e.g., Hausgarten, Ocean Networks Canada).

We need distributed observations of environmental change in the deep ocean, faunal responses to this change and the consequences for people (ecosystem services) to better understand the deep ocean and predict the earth climate system. This will guide how we can mitigate and adapt to change, and perhaps introduce a new set of blue carbon solutions (Hilmi et al., 2021). But this is not a main message of the IPCC SROCC, and there is very little about the deep ocean in the summary for policy makers (IPCC, 2019). Much assembled material addressing climate impacts on deep-ocean ecosystems and consequences for ecosystem services was cut or merged, and in a few cases relegated to supplemental material to meet page limits and emphasize known risk in the SROCC (Bindoff et al., 2019). It is probable that similar cuts have occurred for other ecosystems in SROCC and other reports, but it is doubtful that any of those systems cover an area as vast or are as poorly known as the deep ocean.

DISCUSSION

The IPCC may need to begin producing parallel *inverse* assessment reports that shine a light on what we don't know but need to know and why, to help guide the efforts of science, scientists and science funders in the coming decade. Such an inverse report might identify the deep-sea processes that need to be understood and parameters measured to improve (or groundtruth) climate model predictions and the geographic regions and ecosystems where direct observations would enhance understanding of these processes. It could also highlight additional deep-sea research, technological advances and modeling innovations required for management of resource extraction (e.g., for seabed mining) and biodiversity conservation. Notably, the deep-sea ecosystem subsections of the World Ocean Assessment II (United Nations, 2021) devote considerable attention to knowledge gaps relevant to societal needs. The IPCC could draw on this massive compilation and synthesis to a greater extent.

There is growing attention to the ocean-climate-biodiversity nexus among scientists, but it needs to deepen. The deep sea rarely appears in discussion of integrated strategies to achieve ocean-related climate and biodiversity action synergies. Understanding deep-sea species tolerances and thresholds to climate variables, traits that exacerbate climate vulnerability, the consequences of changing species distributions and interactions, how these translate into altered ecosystem function and the implications for the ecosystem services we rely on, is fundamental. This knowledge can underpin climate adaptation and carbon conservation efforts that maintain carbon sequestration and protect the largely undescribed biodiversity in the face of rising threats from bottom trawling, deep oil and gas extraction and seabed mining. It also may eventually be needed to understand the consequences of different forms of climate intervention or geoengineering. By informing IPCC assessments, UNFCCC Nationally Determined Contributions, Research Dialogues, and the Global Stocktake, deep-sea science can help achieve the goals of the UNFCCC Paris Agreement.

The UN Decade for Ocean Science offers an important opportunity to fill deep-ocean knowledge gaps. A Decade deep sea "community of practice," with programs such as the Deep-Ocean Observing Strategy (Levin et al., 2019) and Challenger 150 (Howell et al., 2021), could generate observations and build knowledge needed for sustainable development of the ocean and people under climate change. Such deep-ocean knowledge might make tractable questions such as: How can natural carbon processes be harnessed to help ameliorate or mitigate climate change? and How can we best manage human activities and emerging industries to maintain these processes? A deeper understanding of the deep ocean will ultimately define what we exploit and what we protect.

DATA AVAILABILITY STATEMENT

The original contributions presented in the study are included in the article/supplementary material, further inquiries can be directed to the corresponding author/s.

AUTHOR CONTRIBUTIONS

LL originated and wrote the manuscript.

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Internal Ocean Dynamics Control the Long-Term Evolution of Weddell Sea Polynya Activity

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Open-ocean polynyas effectively couple the ocean and atmosphere through large ice-free areas within the sea-ice cover, release vast quantities of oceanic heat, and impact deep ocean ventilation. Changes in polynya activity, particularly in the Weddell Sea, may be key to longer time-scale climate fluctuations, feedbacks and abrupt change. While changes in the occurrence of Weddell Sea polynyas are generally attributed to changes in the atmospheric surface forcing, the role of internal ocean dynamics for polynya variability is not well-resolved. In this study we employ a global coupled ocean-sea ice model with a repeating annual atmospheric cycle to explore changes in Weddell Sea water mass properties, stratification and ocean circulation driven by open-ocean polynyas. During the 1300-year long simulation, two large polynyas occur in the central Weddell Sea. Our results suggest that Weddell polynyas may be triggered without inter-annual changes in the atmospheric forcing. This highlights the role of ocean processes in preconditioning and triggering open-ocean polynyas on multi-centennial time-scales. The simulated polynyas form due to internal ocean-sea ice dynamics associated with a slow build-up and subsequent release of subsurface heat. A strong stratification and weak vertical mixing is necessary for building the subsurface heat reservoir. Once the water column turns unstable, enhanced vertical mixing of warm and saline waters into the surface layer causes efficient sea ice melt and the polynya appears. Subsequent, vigorous deep convection is maintained through upwelling of warm deep water leading to enhanced bottom water formation. We find a cessation of simulated deep convection and polynya activity due to long-term cooling and freshening of the subsurface heat reservoir. As subsurface waters in the Southern Ocean are now becoming warmer and saltier, we speculate that larger and more persistent Weddell polynyas could become more frequent in the future.

Keywords: Weddell polynya, ocean mixing, climate model, sea ice, deep water formation, Southern Ocean, internal ocean dynamics

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1. INTRODUCTION

The Southern Ocean is an important region for formation of cold and dense bottom waters feeding the global ocean circulation and plays a key role in regulating Earth's climate by redistributing heat and regulating the uptake of atmospheric CO_2 (Orsi et al., 1999; Marshall and Speer, 2012; Watson et al., 2014). In the Weddell Sea, bottom water is formed on the shallow continental shelf in small coastal polynyas, as sea ice is pushed away from the Antarctic continent by strong winds, exposing the relatively warm ocean to the cold polar atmosphere. Intense sea ice formation and brine rejection leads to the formation of high salinity shelf water that spills over and forms Antarctic Bottom Water (AABW) (Orsi et al., 1999). This process, however, is poorly represented in today's climate models with the majority of the models contributing to the 5th phase of the Coupled Model Intercomparison Project (CMIP5) forming bottom water by open-ocean deep convection in the Weddell Sea (Heuze et al., 2013; Heywood et al., 2014). This often leads to significant biases in modeled Southern Ocean stratification affecting deep ocean volume transports (Timmermann and Beckmann, 2004; Behrens, 2016). This highlights the importance of understanding the processes leading to deep convection in climate models in order to improve future projections (Zhang et al., 2019).

In the models, deep convection is closely linked to the formation of large open-ocean (or sensible heat) polynyas in the Weddell Sea. As opposed to coastal polynyas, Weddell polynyas form off-shore in response to upwelling of Warm Deep Water (WDW) located below a weakly stratified surface layer (e.g., Martin et al., 2013; Cheon et al., 2015). In the winter, when the stratification is weak, small changes in surface buoyancy can trigger deep convection causing WDW to be mixed into the surface layer where it melts sea ice, forming a region of open water within the sea ice pack. Intense cooling at the air-sea interface forces surface water to cool and sink in the polynya, presenting an alternative source of deep and bottom waters (Killworth, 1983). Consequently, Weddell polynyas have widespread implications for deep ocean ventilation and uptake of anthropogenic and natural CO_2 (Bernardello et al., 2014; Menviel et al., 2018), overturning circulation strength (Martin et al., 2013, 2015), and impact Antarctic as well as global climate variability (Pedro et al., 2016; Cabré et al., 2017; Zhang et al., 2019). Recently, Naughten et al. (2019) also showed that Weddell Sea polynyas may accelerate melting of nearby Antarctic ice shelves by increasing the transport of WDW into the ice shelf cavity. Polynyas have also been invoked as a potential mechanism to explain rapid climate fluctuations in the North Atlantic during the last glacial period (Vettoretti and Peltier, 2016).

In contrast to the CMIP5 models, exhibiting frequent open-ocean deep convection, the occurrence of large polynyas is a rare phenomenon in the instrumental record. During the mid-1970s, satellite observations revealed a large (250,000–300,000 km^2) winter-persistent polynya in the eastern Weddell Sea (Carsey, 1980). This feature, which is known as the Great Weddell Polynya, persisted for three consecutive winters (1974–1976) and formed after a prolonged period of cold and dry atmospheric conditions. This contributed to a saltier surface layer and a weakening of the surface stratification enabling WDW to reach the sea ice (Gordon et al., 2007). Several smaller and intermittent polynyas were observed during the 1990s over Maud Rise (Smedsrud, 2005), and again in 2016 and 2017 reaching a size of 50,000 km^2 (Swart et al., 2018; Campbell et al., 2019; Jena et al., 2019).

Unfortunately, due to the short instrumental record, there are large uncertainties in the long-term evolution of Southern Ocean polynya activity. Although recent attempts have been made to reconstruct past polynya activity, the results are still ambiguous (Goosse et al., 2021), and the climatic conditions necessary to sustain deep convection in the Southern Ocean are not well constrained. de Lavergne et al. (2014) speculated that open-ocean convection in polynyas may have presented a dominant mode of deep water formation in the past, but has reduced over the past couple of decades due to a freshening of the Southern Ocean surface waters. Similarly, other studies point to large-scale changes in atmospheric circulation patterns as being the controlling factor for polynya formation and its absence in recent decades (e.g., Gordon et al., 2007; Cheon et al., 2017; Campbell et al., 2019; Kaufman et al., 2020). Meanwhile, less attention has been given to how internal ocean-sea ice processes might influence stratification (especially over long timescales) and its role in preconditioning and potentially triggering open-ocean polynyas. In particular, given its importance for polynya formation, changes in WDW heat content (e.g., by mixing processes in the ocean) may present a major control on the frequency of polynya events. In climate models, the simulated open-ocean convection is often preceded by a build-up of heat and salt at mid-depths over several years or decades (Martin et al., 2013; Cheon et al., 2015; Zanowski et al., 2015; Dufour et al., 2017). The build-up eventually destabilizes the water column, triggering cycles of deep convection and subsequent build-up. This is also evident in observations by Robertson et al. (2002) and Smedsrud (2005), showing that the WDW warmed from the late 1970s to the 1990s when the Weddell Polynya was absent. Moreover, it has speculated that the observed warming of Southern Ocean deep waters and shrinking of AABW in the last decades may reflect changes in polynya activity (e.g., Purkey et al., 2012; Zanowski et al., 2015; Wang et al., 2016). While it remains debated whether subsurface heat accumulation, which is simulated in models, can trigger polynyas (e.g., Campbell et al., 2019), it nevertheless suggests a close connection between the WDW heat content and the occurrence of Weddell polynyas.

In this study, we explore the impact of long-term changes in WDW and Weddell Sea stratification and how it affects the formation of open-ocean polynyas. Applying a stand-alone ocean-sea ice model with a constant atmospheric forcing, we analyse a 1300-year transient simulation where two large Weddell polynyas form. This presents a unique opportunity to test how interior ocean processes influence the frequency of Southern Ocean deep convection events in the absence of atmospheric variability. Rather than trying to explain the trigger mechanism for the observed Weddell Polynya in the 1970s, this study focuses on the general dynamics of polynya formation by addressing two main questions: (1) Can build-up of subsurface heat and salt alone trigger Weddell polynyas? and (2) What controls the time-scale and frequency of polynya events in the Southern Ocean?

2. METHODS

2.1. Model Description

To study the polynya dynamics, we use a stand-alone ocean-sea ice configuration of the Norwegian Earth System Model (NorESM-OC1.2; Schwinger et al., 2016). The ocean-sea ice component of this NorESM-OC1.2 is the same as in Bentsen et al. (2013), and there is also a carbon-cycle part which has not been used in this study. Changes in the carbon-cycle induced by Southern Ocean convection is beyond the scope of this paper, but could be relevant to explore in future research.

The family of NorESM models is based on the Community Climate System Model version 4 (CCSM4), using the same land and sea-ice models, but with a different atmosphere and ocean component. The ocean model is based on the Miami Isopycnal Coordinate Ocean Model (MICOM). MICOM uses potential density as the vertical coordinate which differs significantly from z-coordinate models by mimicking the real structure of the ocean as discrete layers of constant density. One of the key advantages of the isopycnal coordinate models is reduced spurious mixing due to the numerical representation of isopycnal advection and mixing. This allows for a more accurate transport, improved representation of dense water overflows in gravity currents and preservation of water mass characteristics (Griffies et al., 2000; Assmann et al., 2010; Bentsen et al., 2013). The sea ice model is the Community Ice Code version 4 (CICE4) which is the same sea-ice component used in CCSM4 (Hunke and Lipscomb, 2008; Gent et al., 2011). CICE4 is a fully dynamic-thermodynamic sea ice model and shares the same horizontal grid as the ocean component.

The model is configured on a tripolar grid with a longitudinal resolution of 2° and a variable latitudinal grid spacing from 0.5° at the Equator to 0.35° at high southern latitudes. In the vertical, the ocean is divided into 51 isopycnal layers with potential density classes ranging from 23.602 to 33.2 kg m^{-3}. The first two model layers constitute the mixed layer where the density can evolve freely. The uppermost layer is limited to 10 m if the total mixed layer depth is <20 m, thus allowing the surface ocean to respond faster to surface fluxes. The mixed layer depth is parameterized according to the turbulent kinetic energy (TKE) model based on Oberhuber (1993). To reduce SSS biases in high latitude regions, salt released due to sea ice formation is evenly distributed below the mixed layer down to a depth where the density difference is 0.4 kg m^{-3} relative to the surface. This avoids build-up of salt and improves stratification in seasonally ice-covered oceans (Bentsen et al., 2013).

In isopycnal models potential density is referenced to a given pressure. This means that the flow characteristics are most accurately represented near the reference level, where isopycnal and neutral density surfaces are similar. When the pressure deviates significantly from the reference pressure, i.e., in the deep ocean, the potential density becomes increasingly non-neutral. In order to avoid this non-neutrality, most isopycnal models now use a potential density referenced to 2,000 db, which maximizes the neutrality of isopycnal layers (McDougall et al., 2005). In high-latitude regions, where the water column is weakly stratified, the stratification is more sensitive to the choice of the reference pressure. In a previous version of NorESM, Assmann et al. (2010) showed that the 2,000 db reference pressure caused the stratification in the Weddell Sea to be unstable, leading to overestimated vertical mixing which eroded the WDW. As the subsurface heat reservoir is essential for the dynamics of polynya formation and deep convection in the Weddell Sea (Martin et al., 2013; Gordon, 2014; Cheon et al., 2015; Dufour et al., 2017) we chose a reference pressure of 1,000 db. This ensures a stable stratification in the Weddell Sea region and improves the representation of the WDW layer. We note, however, that this could introduce biases in the deepest part of the basin where the pressure can differ significantly from the potential density at 1,000 db.

The isopycnal formulation used offers complete control of the diapycnal mixing applied in the model. The total diffusivity of diapycnal mixing consists of four components relating to different mixing processes in the interior ocean:

$$\kappa = \kappa_b + \kappa_r + \kappa_t + \kappa_N \tag{1}$$

$$Ri_g = \frac{N^2}{(\delta u/\delta z)^2 + (\delta v/\delta z)^2}, \quad N^2 = \frac{g}{\rho}\frac{\delta\rho}{\delta z} \tag{2}$$

where κ_b is a constant background diapycnal diffusivity set to 10^{-5} m^2 s^{-1}, including a latitude dependency according to Gregg et al. (2003). κ_r is the shear driven diapycnal mixing as a function of the local gradient Richardson number which relates to the balance between shear induced turbulence and buoyancy forcing. The tidally driven mixing κ_t induced by the internal wave field and braking of internal waves is implemented from Simmons et al. (2004). Lastly, κ_N relates to the diapycnal mixing when local stability is weak and is triggered if the local interface density difference is <0.006 kg m^{-3}. A convective adjustment is then made at each time step by increasing the diapycnal diffusivity to 0.05 m^2 s^{-1}. Mixing along isopycnal surfaces is parameterized according to Eden and Greatbatch (2008).

A more detailed description of the NorESM ocean component can be found in Assmann et al. (2010), Bentsen et al. (2013), and Schwinger et al. (2016).

2.2. Experimental Setup

The model is forced with the Coordinated Ocean-ice Reference Experiment version 1 (CORE-I) repeated normal year forcing (CORE-NYF) which has been derived from 43 years of inter-annually varying forcing with a seamless transition from 31 December to 1 January (Large and Yeager, 2004). The CORE framework is used as a tool to study the behavior in coupled global ocean and sea ice models forced by a common atmospheric dataset (Griffies et al., 2009). This reduces the complexity of the coupled climate system by prescribing atmospheric boundary conditions without dealing with the ocean-atmosphere coupling and feedbacks. The normal year forcing is constructed by repeating a 1-year annual cycle, which means that there is no variability in the atmospheric forcing beyond seasonal variations. Thus, the inter-annual variability simulated in the coupled ocean-sea ice model is due to internal dynamics. This approach allows us to explore the physical processes internal to the ocean-sea ice system in response to a repeated annual atmospheric cycle and test if Weddell Sea polynyas can form without invoking changes to the atmospheric forcing. The model is initialized with January mean temperature and salinity fields from the Polar Science Center Hydrographic Climatology (Steele et al., 2001), and the model is integrated for 1300 years.

There is a rapid adjustment during the first ~50 years of the simulation, during which the simulated Southern Ocean sea ice extent and upper ocean properties are equilibrating to the surface forcing. After the initial adjustment period, the Weddell Sea water mass properties undergoes a long-term adjustment over the 1300-year simulation period (see **Supplementary Figure 1** for the full time series), which is related to the occurrence of two large polynyas forming in the central Weddell Sea. The first polynya event occurs after year 133 in the simulation, persists for 25 years, and is followed by a long period of 300 years without deep convection. The second polynya occurs from year 472 to 500 with approximately the same size and duration as the first event. In the remaining part of the simulation no polynya forms. Hence, this simulation allows us to investigate how long-term changes in Weddell Sea stratification and WDW heat content influence the formation of Weddell polynyas, and identify key processes leading to the cessation of open-ocean deep convection.

3. RESULTS

The first polynya initially develops in the deep part of the central Weddell Sea at 23–5°W and 60–65°S, hereon referred to as the polynya region (orange box in **Figure 1**). This region is characterized by a doming of isopycnals, causing a weakening of the local stratification, and possibly promoting the formation of polynyas (Gordon et al., 2007).

In order to diagnose the processes leading to the formation of the polynya in the NorESM-OC1.2 simulation, we limit our analysis to the first 800 years of the simulation. However, the first 50 years are excluded, to avoid any effects associated with the initial model adjustment. First, a detailed analysis of the first polynya event is performed, where we focus on three main time periods: (1) A preconditioning phase (year 50–100) where heat and salt builds up in the WDW layer; (2) A mixing phase (year 100–133) where the WDW starts to erode and sea ice concentration decreases; and (3) A convective phase (year 133–160) marked by the opening of a small polynya which triggers open-ocean deep convection and forcing an abrupt and extensive retreat of sea ice in the Weddell Sea. The sea ice conditions corresponding to each of the different phases are illustrated in **Figure 2**. After analysing the first polynya event we compare it to the second event, assessing which processes are critical for the polynya to reoccur.

A comparison between the Levitus climatology (Levitus, 1983) and simulated temperature and salinity fields in the Southern Ocean is shown in **Supplementary Figure 2**. The simulated water mass properties agrees reasonably well with the observational data showing a northward intrusion of low salinity Antarctic Intermediate Water and a southward extension of relatively warm and salty North Atlantic Deep Water (NADW). In the interior Weddell Sea, the WDW is observed as a distinct temperature maximum at mid-depth and originates from entrainment of relatively warm Circumpolar Deep Water (CDW) by the Weddell Gyre (Orsi et al., 1993). However, the simulated WDW layer sits deeper in the water column compared to the observations (see also Robertson et al., 2002; Fahrbach et al., 2004). We attribute this to the relatively weak AABW formation in the model, when deep convection is absent thus resulting in a downward displacement of deep isopycnals causing the core of the WDW to migrate downwards. The implications of the deep WDW layer for the dynamics of the simulated polynya is discussed further in section 4.1. Previous studies have

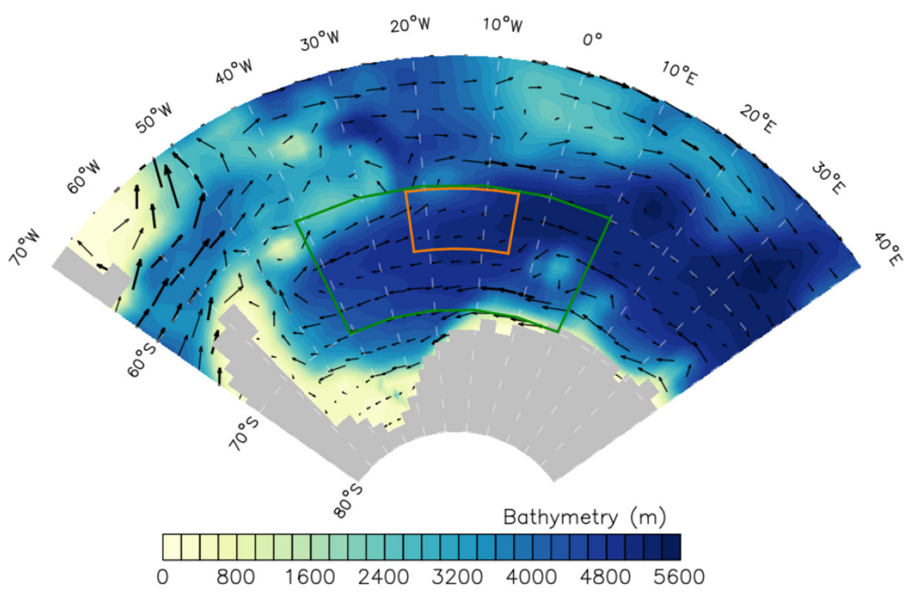

FIGURE 1 | An overview of bathymetry (m) and circulation in the Weddell Sea. The orange box at 23–5°W and 60–65°S shows the region where the first polynya initially forms. The Weddell Sea Convective Region (WSCR) (green box) is defined as the area from 31°W to 9°E and 60 to 70°S. Black arrows show the barotropic velocity field with thicker arrows corresponding to higher velocities.

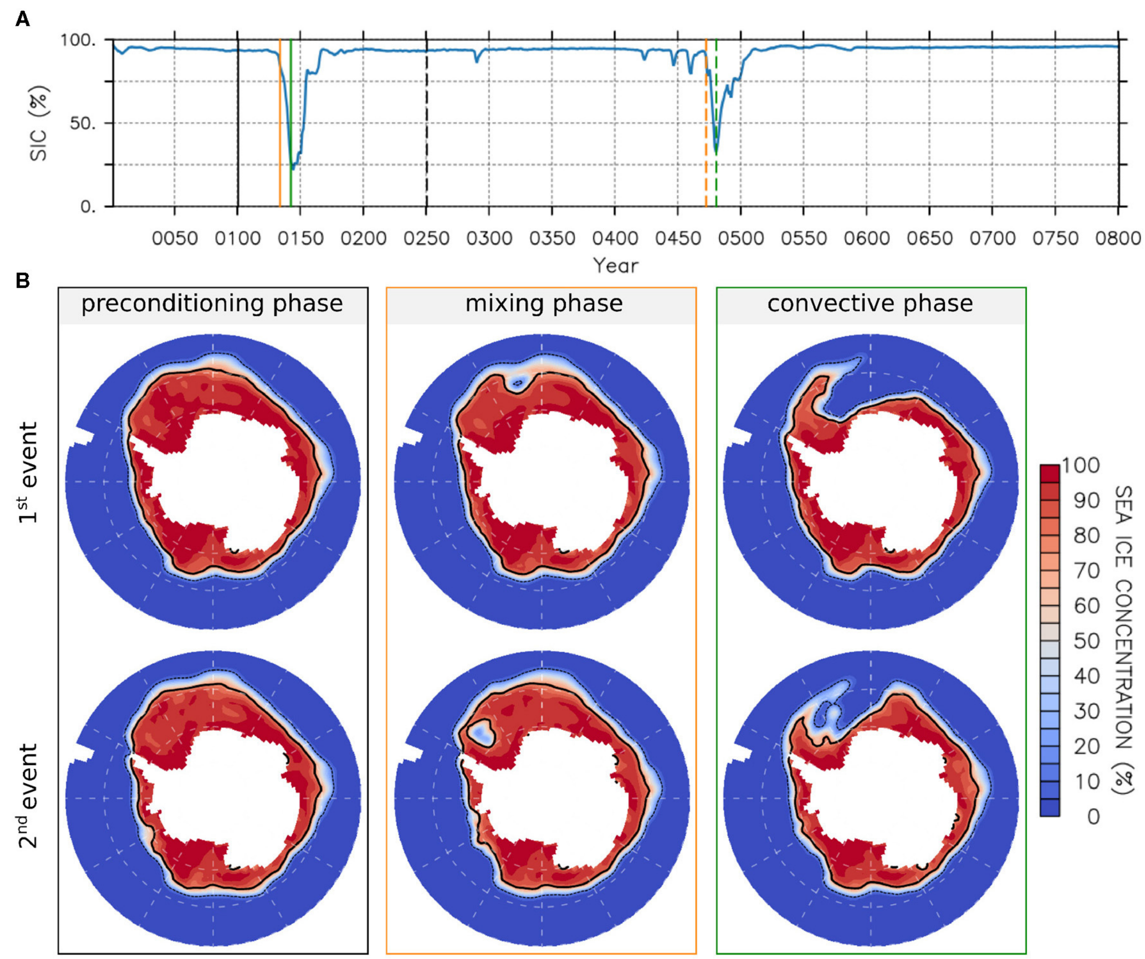

FIGURE 2 | Winter mean (July-August-September) sea ice concentration (%) in the Southern Ocean. **(A)** Time series of sea ice concentration averaged over the green box in **Figure 1**, with vertical lines marking the end of the preconditioning (black), mixing (orange), and convective (green) phases for the first (solid) and second polynya event (dashed). **(B)** Spatial extent of sea ice concentration for the first and second polynya at the end of the preconditioning phase (year 100; year 250), the onset of the polynya (year 133; year 472) and convective phase (year 142; year 480). The solid and dashed black lines mark the 75 and 15% sea ice concentration, respectively.

shown that general circulation models struggle with correctly representing water mass properties in the Southern Ocean and therefore exhibit significant biases relative to observations (e.g., Heuze et al., 2013; Downes et al., 2015). Similarly, in the Arctic, models commonly produce an Atlantic Water layer that is too thick and too deep (e.g., Ilicak et al., 2016). While the simulated stratification in the Weddell Sea region differs somewhat from observations, our model captures the main features of the water mass structure, which allows us to explore the connection between open-ocean polynyas and long-term changes in WDW properties.

3.1. Phase I—Preconditioning Phase

The preconditioning phase (year 50–100), corresponds to a period where no polynyas occur. The simulated winter mean sea ice concentration (**Figure 2A**) and thickness (not shown) in the Weddell Sea is relatively stable throughout the preconditioning phase and compares reasonably well with observations (Parkinson and Comiso, 2008). Thick and compact multi-year sea ice is generally found along the continental shelf and the Antarctic Peninsula due to strong sea ice formation and convergence in the sea ice drift. In the central part of the Weddell Sea, where the polynya initially forms, the seasonal sea ice cover is typically less thick (about 0.6 m) and maximum sea ice concentration ranges between 90 and 100%. The stable winter sea ice cover in the polynya region is sustained by a relatively weak density gradient—the pycnocline—separating warm and salty water in the WDW layer from the surface (**Figure 3C**). The upper 200 m, corresponding to the depth of the winter mixed layer, is relative cold and fresh due to freshwater input from seasonal sea ice melt. Below the winter mixed layer, we find the relatively homogeneous and weakly stratified Modified Warm Deep Water (MWDW) which refers to WDW that has been mixed with colder surface waters. The upper limit of WDW, at

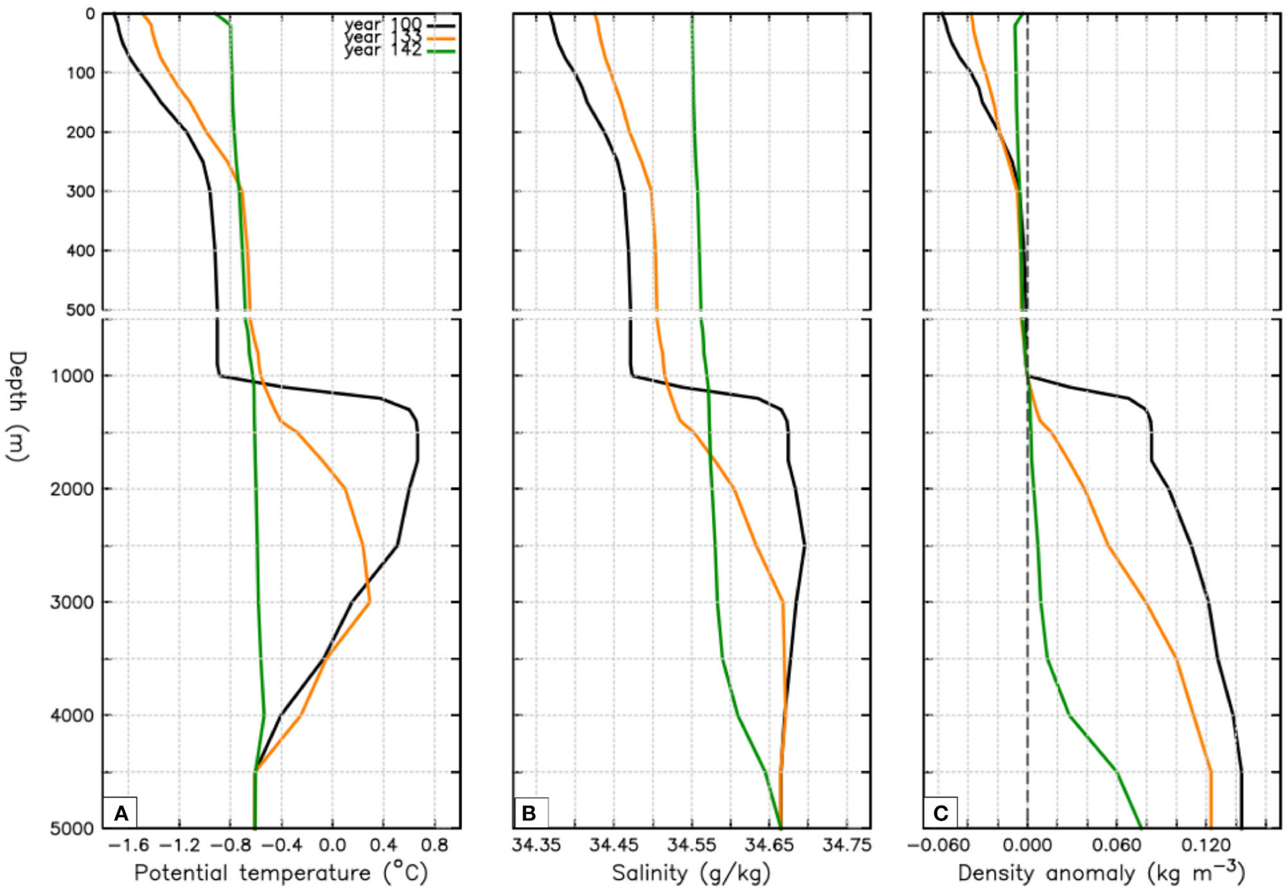

FIGURE 3 | Winter (September) vertical profiles averaged over the polynya region (orange box in **Figure 1**) for years 100 (black), 133 (orange), and 142 (green) of **(A)** potential temperature (°C), **(B)** salinity (g/kg), and **(C)** potential density (σ_0) anomalies with respect to the density at 1,000 m. Density is calculated using the 1980 UNESCO International Equation of State (IES80) and referenced to 0 dbar.

1,000 m depth, is characterized by temperatures >0°C and is separated from the MWDW by a strong gradient in temperature and salinity. Vertical profiles of potential density in **Figure 3C** illustrate that stratification from the surface down to 2,500 m is strongly dominated by salinity, while decreasing temperatures below the WDW control deep ocean stratification.

During the preconditioning phase we find a gradual build-up of heat and salt in the WDW layer (**Figures 4C,D**). The accumulation of heat and salt at mid-depth is enabled by a stable stratification, that persists throughout the preconditioning phase and isolates the WDW from surface interactions. At the end of the preconditioning phase (year 100), the core of the WDW layer is located between 1,000 and 2,000 m depth, with a temperature maximum ($\theta_{\max}$) of 0.64°C and salinity 34.67g kg^{-1}. This corresponds to a mean warming rate of about 0.005°C yr^{-1} and 0.009 g kg^{-1} yr^{-1} increase in salinity over the 50 year period. This is within the uncertainties of the observational estimates from Robertson et al. (2002) of 0.012 ± 0.007°C yr^{-1} warming of WDW following the 1970s Weddell Polynya. Interestingly, $\theta_{\max}$ peaks 30 years before the polynya opens, indicating that the build-up of subsurface heat alone is not enough to trigger deep convection. Rather, the heat reservoir preconditions polynya formation by gradually weakening stratification and fuels deep convection once the polynya has formed.

3.2. Phase II—Mixing Phase

From year 100, $\theta_{\max}$ starts to decrease, which marks the onset of the mixing phase. Above the WDW layer, temperature and salinity increases (**Figures 3A,B**), and the upper ocean stratification weakens (**Figure 3C**), leading to reduced sea ice concentration in the polynya region. However, the polynya does not appear before year 133 (**Supplementary Figure 3**).

To understand the processes leading to the opening of the polynya, we look in detail on the time period from year 100 to 150 in **Figure 5**. The average sea ice thickness in the polynya region is 0.6 m in winter and melts away entirely in the summer (**Figure 5A**), maintaining a strong surface stratification and facilitating sea ice growth the following winter. Starting in year 125, 8 years prior to the opening of the polynya, sea ice thickness gradually decreases. The reduced sea ice thickness can be linked to an increase in bottom-up fluxes of heat and is suggestive of a vertical diffusion of heat from the warmer

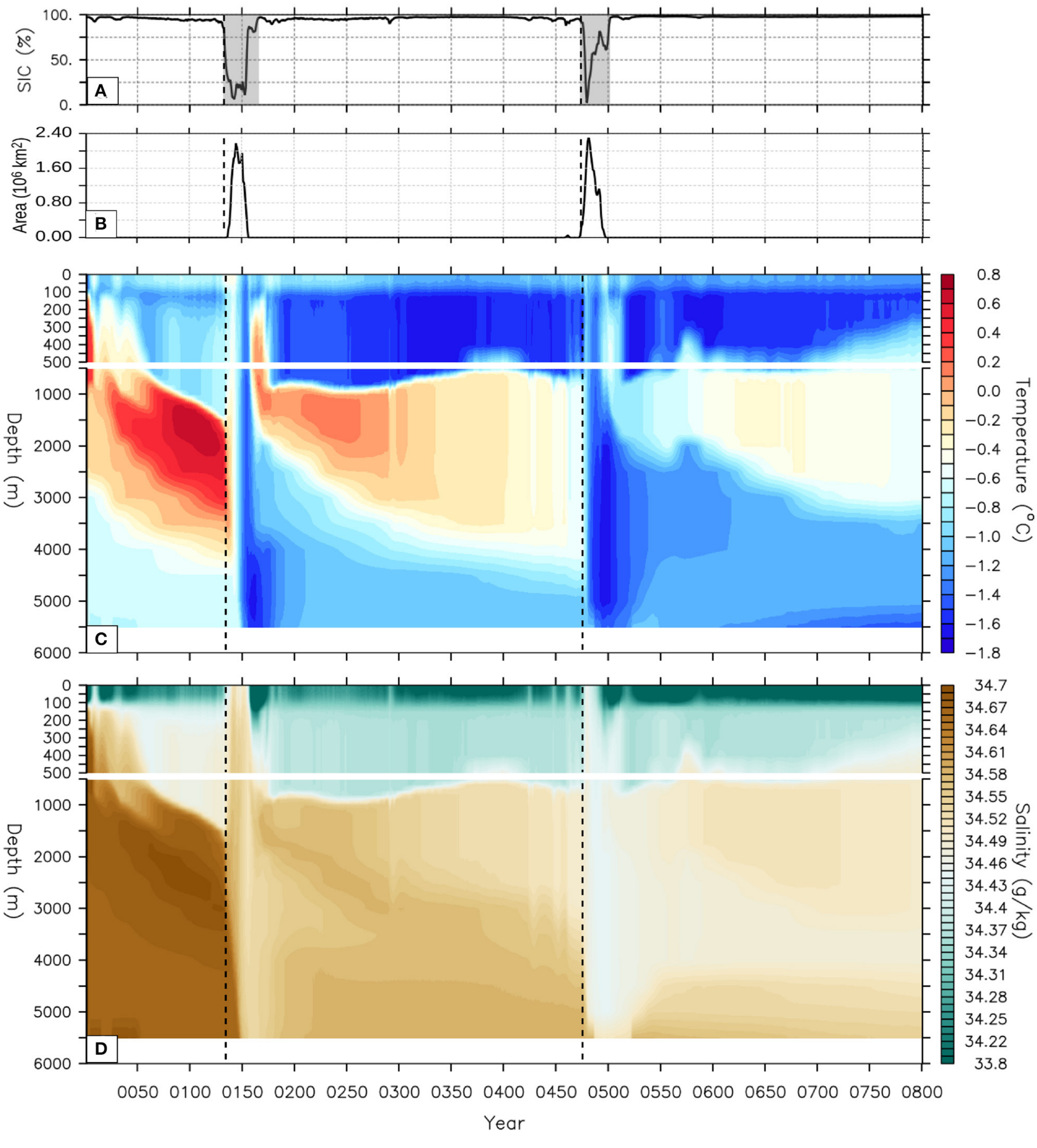

FIGURE 4 | Time series of **(A)** September sea ice concentration (%), **(B)** total convective area (10^6 km^2) and vertical profiles of annual mean **(C)** potential temperature (°C) and **(D)** salinity (g/kg) averaged over the polynya region (23–5°W, 60–65°S) for the first 800 years of the simulation. The convective area is defined as the integrated area where the annual maximum mixed layer depth exceeds 2,000 m. In **(C,D)** the upper 500 m has been exaggerated for a better view of the surface stratification. Black dashed lines mark the opening of the first and second polynya and gray shaded areas show periods of active deep convection.

subsurface layer into the mixed layer. This heat flux melts the sea ice, while maintaining the winter SST at the freezing point (**Figure 5B**). When the winter sea ice thickness drops below 0.5 m, the ocean-atmosphere heat flux increases (**Figure 5D**), thus stimulating further thinning of the sea ice. At the end of the mixing phase, the winter-mean sea ice thickness has decreased to 0.3 m. The annual mean freshwater flux (expressed in terms of sea ice equivalents) associated with melting and freezing of sea

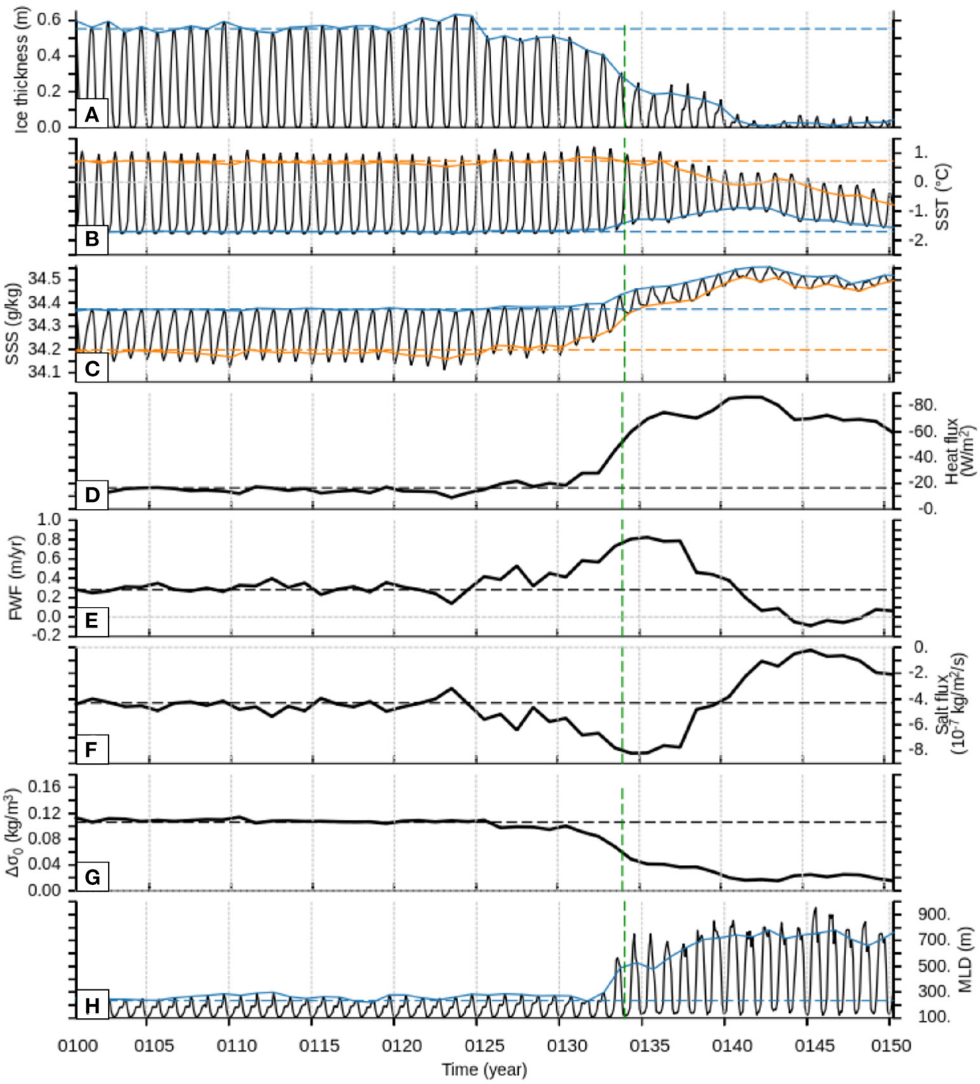

FIGURE 5 | Time series of model variables averaged over the polynya region (orange box in **Figure 1**) during the mixing phase (year 100–150). **(A)** Sea ice thickness (m), **(B)** SST (°C), **(C)** SSS (g/kg), **(D)** annual mean atmosphere-ocean heat flux (W m^{-2}), **(E)** annual mean fresh water flux due to melting and freezing of sea ice (m yr^{-1}). The freshwater flux is converted into sea ice thickness equivalents over one year assuming an average sea ice density of 900 kg m^{-3}. Positive (negative) values are associated with sea ice melt (formation). **(F)** Total salt flux (kg m^{-2} s^{-1}) received by the ocean at the surface (positive values increases ocean salinity). **(G)** The density stratification ($\Delta\sigma_0$; kg m^{-3}) calculated as differences between 100–200 and 0–100 m. **(H)** Maximum mixed layer depth (m) from the model. Blue and orange lines corresponds to winter (September) and summer (March) and dashed lines shows the mean pre-polynya values. The dashed green line at year 133 marks the onset of the convective phase.

ice (**Figure 5E**) increases in parallel with the sea ice thinning and increasing heat fluxes. Early in the mixing phase there is a net positive surface freshwater flux, equivalent to 0.3 m of sea ice melt over a year, indicating that more sea ice melts than what is formed locally in the polynya region. Hence, sea ice, that is formed non-locally, is advected into the region where it melts and freshens the surface layer. As the opening of the polynya approaches, the freshwater flux increases due to a combination of reduced sea ice formation in winter (less brine rejection) and increased advection of sea ice into the region. This tends to stabilize the water column, providing a negative feedback that oppresses further sea ice melt and prevents deep convection (Martin et al., 2013). However, the fact that sea ice thickness continues to decrease shows that the freshwater input is too weak to suppress the strong heat flux from below.

Despite of the increased surface freshwater flux from sea ice melt, there is an increase in the surface salinity prior to the opening of the polynya (**Figure 5C**). This is explained by the enhanced upward fluxes of warm, saltier water from below, rather than changes in surface fluxes. To demonstrate this, the total salt flux is plotted in **Figure 5F** and shows contributions from the atmosphere and sea ice model to the surface salinity (i.e., precipitation-evaporation, liquid and frozen run-off, and thermodynamic sea ice growth/decay), but excludes salt fluxes from the ocean model itself. Because the total surface salt flux is negative throughout the mixing phase (mainly due to sea ice melt; see **Figure 5E**), this indicates that the increase in SSS must be due to an oceanic salt flux into the mixed layer from below and/or horizontally. This preconditions the opening of the polynya in year 133 by weakening the upper ocean stratification (**Figure 5G**). The mixed layer depth (**Figure 5H**) does not change until after the polynya emerges. This supports the fact that the small polynya opening is not triggered by deep convection (through surface buoyancy or wind-stress forcing). Rather, it is only after the polynya has been established that deep convection can occur (Gordon, 1982).

3.3. Phase III—Convective Phase

In year 133 a small opening in the sea ice cover appears at 23–5°W, 60–65°S during mid-winter (**Figure 2**; center middle panel). The opening of the small polynya triggers open-ocean deep convection by exposing the underlying ocean to the cold atmosphere, abruptly increasing the ocean-atmosphere heat flux by close to 100 W m^{-2}. Intense cooling at the air-sea interface leads to enhanced surface buoyancy loss and the water column becomes unstable, triggering deep convection, and mixing over the whole water column (**Figure 3C**). As the winter mixed layer deepens (**Figure 5H**), WDW is injected directly into the surface layer. This leads to winter SSTs well above the surface freezing point (**Figure 5B**), thus preventing ice formation. A few years after the polynya appears, a large embayment has formed in the Weddell Sea (**Figure 2B**; upper right panel).

For an indication of the strength of the simulated deep convection and polynya extent, we calculate total area of convection by summing-up the total number of grid cells where the MLD exceeds 2,000 m following de Lavergne et al. (2014) (**Figure 4B**). We use the MLD criterion from Martin et al. (2013), corresponding to the depth where the density differs by 0.01 kg m^{-3} from its surface value. When the polynya opens in year 133, the convective area is only 57,150 km^2, but grows to a maximum of 2,157,000 km^2 in year 142 (**Figure 4B**). In comparison, the 1970's Weddell Polynya was about 300,000 km^2. This is a common bias in fully-coupled, coarse resolution climate models, producing excessive deep convection and overestimating polynya size (averaging 930,000 km^2 in 70% of pre-industrial CMIP5 model simulations; de Lavergne et al., 2014). We attribute the larger simulated polynya to the substantial ocean heat loss during the convective phase. The change in ocean heat content is shown in **Figure 6**, and reflects the WDW cooling and diminishing WDW layer thickness. Heat depletion is not only confined to the convection region, but affects the entire Atlantic sector with a heat loss of more than 14 GJ m^{-2} in the central Weddell Sea. This corresponds to a cooling of WDW by 1.8°C (θ_{max} drops from 0.6 to −1.2°C) by the end of the convective phase. The cooling is faster than the resupply of heat from the inflow of CDW, which ceases during the convective phase. In **Figure 6A** this can be seen as a strong positive anomaly in the South Atlantic corresponding to an accumulation of subsurface heat there. Convection stops in year 160 after the heat reservoir is depleted. Without the upwelling of warm water, the sea ice cover regrows and a cold and fresh surface layer is quickly established thus allowing heat and salt to build up again (**Figure 4**).

Deep convection associated with the polynya triggers a large increase in bottom water formation (Martinson et al., 1981). This is illustrated in **Figure 7**, showing the age of water (AOW) at different depths, which is used as a diagnostic for ocean ventilation. A few years before the polynya opens the AOW at 2,000 m decreases, while it increases above the WDW layer at 500 m depth. This points to an increased vertical exchange between the WDW and the more ventilated water mass above. Once the polynya opens we find young water at 4,000 m depth, while near-surface waters are getting older as it is mixed with less ventilated deep water. This a clear indication of open-ocean deep convection which mixes cold, dense surface water down to

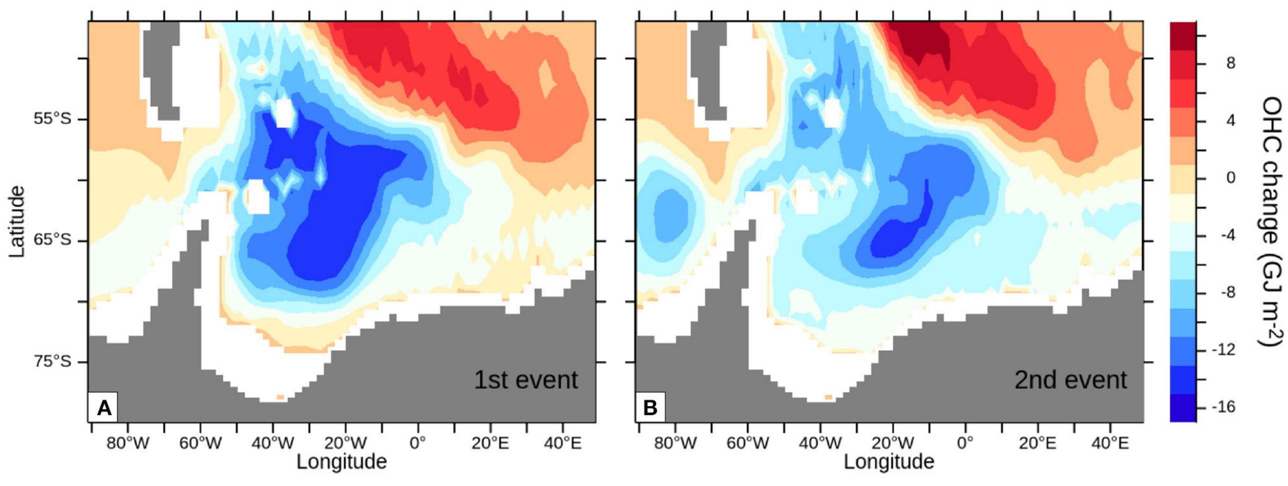

FIGURE 6 | Ocean heat content anomaly (10^9J m^{-2}) between 1,000 and 3,000 m associated with open-ocean deep convection in the Weddell Sea for **(A)** the first polynya and **(B)** second polynya. Anomalies are calculated with respect to the pre-polynya values.

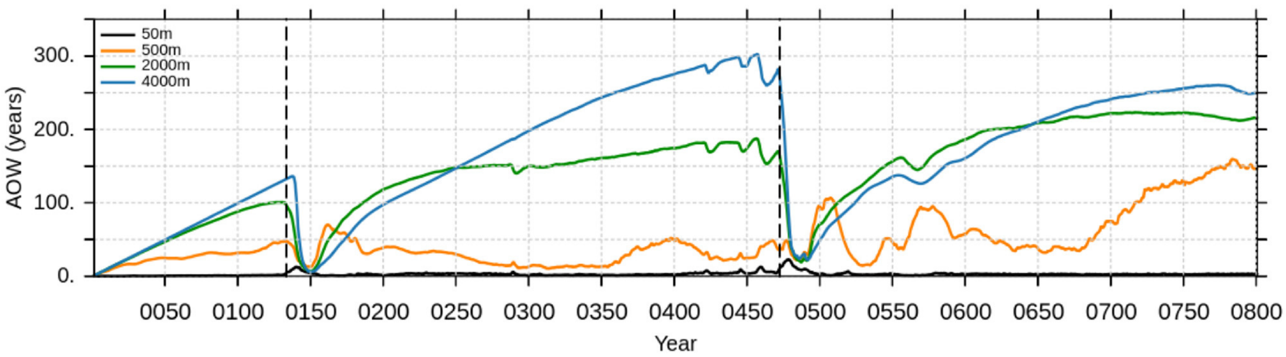

FIGURE 7 | Time series of the Age Of Water (AOW) in years at 50 m (black), 500 m (orange), 2,000 m (green), and 4,000 m depth (blue) in the Weddell Sea Convective Region (41°W – 9°E, 60–70°S) outlined by the green box in **Figure 1**. When the water is at the surface AOW is reset to zero and older ages are thereby associated with less ventilated water. Vertical dashed black lines shows the opening of the two polynyas in year 133 and 472, respectively.

the abyssal ocean. Consequently, the Weddell Sea Bottom Water (below 4,000 m) cool and freshen when convection is active, followed by a mean warming of 0.019°C per decade, when the polynya is absent (**Figures 4C,D**). The post-polynya warming in the abyssal Southern Ocean of 0.002–0.019°C per decade found by Zanowski et al. (2015) is of similar magnitude and is comparable to our result. In response to the stronger deep water formation, the overturning circulation in the Atlantic shows an enhanced northward volume transport of AABW across 30°S at 3,000 m depth by 7–9 Sv ~30 years after the polynya occurs (see **Supplementary Figure 4**). This in good agreement with Zanowski et al. (2015) and Zanowski and Hallberg (2017), who show that it takes 18–25 years for AABW volume transport anomalies, associated with Weddell polynyas, to reach the South Atlantic. Martin et al. (2015) further showed that the enhanced AABW transport associated with Southern Ocean deep convection events may weaken the Atlantic Meridional Overturning Circulation (AMOC) on a multi-centennial time scale. Meanwhile, we only find a small AMOC weakening (2 Sv), indicating that the polynya-driven transport changes are mainly confined to the South Atlantic. This also highlights that open-ocean polynyas could be an important mode of Southern Ocean ventilation, and implies that the observed warming and shrinking of AABW in recent decades could potentially be linked to an absence of Southern Ocean deep convection (Purkey et al., 2012; Zanowski et al., 2015).

3.4. Details of the Trigger Mechanism

In section 3.2, we showed that the opening of the polynya is preceded by increased heat and salt fluxes into the mixed layer leading to warmer and saltier surface waters, weaker stratification and gradual thinning of the winter sea ice cover (**Figures 3**, **5**). This points to diffusive mixing processes as a potential driver for the simulated polynya formation. To illustrate this we show the diffusivity of diapycnal mixing in **Figure 8A**. The highest values of diapycnal diffusivity ($\kappa = 10^{-2}$ m^2 s^{-1}) is generally found in the surface boundary layer, where diapycnal diffusivity is governed by the surface forcing conditions, while in the deep ocean mixing is dominated by the weak background diffusivity ($\kappa = 10^{-5}$ m^2 s^{-1}). The strong gradient in mixing between the boundary layer and the interior ocean reflects how stratification acts to prevent warmer subsurface water from entering the surface layer. From year 110 the diapycnal mixing starts to increase at mid-depth, and intensifies with a couple of strong mixing episodes around year 125 and 130. The change in mixing is largest above the WDW and has increased by an order of magnitude from 10^{-5} to 10^{-4} m^2 s^{-1} at the end of the mixing phase (**Figure 8B**). In the boundary layer, mixing starts to increase 15 years after it increases in the subsurface. This suggest that the enhanced mixing is not driven by changes in surface forcing conditions, but rather arises from destabilizing processes in the deep ocean.

The enhanced vertical mixing results in a gradual increase in temperature and salinity above the WDW interface (**Figures 8C,D**), while the heat and salt content of the WDW layer is decreasing (**Figures 4C,D**). This is consistent with the AOW above the WDW layer getting older (**Figure 7**), while water masses at 2,000 m depth are getting younger. The warmer and saltier water is gradually mixed into the surface layer and temperature and salinity start to increase from year 125. When the polynya opens in year 133, diapycnal mixing is increased over the entire water column mimicking open-ocean deep convection.

As the contribution from shear instabilities to the total diffusivity is generally small (e.g., Dufour et al., 2017) and background diffusivity and tidal mixing is constant, the changes in diapycnal mixing at depth is thought to mainly reflect gravitational instabilities arising from vertical variations in stratification. This is due to the inverse relationship between stratification and vertical mixing (i.e., Equation 2): If the water column is strongly stratified, diapycnal mixing is suppressed, while a weak stratification promotes higher vertical mixing rates. Comparing **Figure 8** and **Figure 3** illustrates how the parameterization of diapycnal mixing is linked to stratification. The erosion of the WDW layer early in the mixing phase causes a buoyancy flux across the upper WDW boundary that weakens the density gradient and reduces the pycnocline strength (**Figures 3C**, **5G**). This can be attributed mainly to the upward salt flux from the WDW layer, that dominates

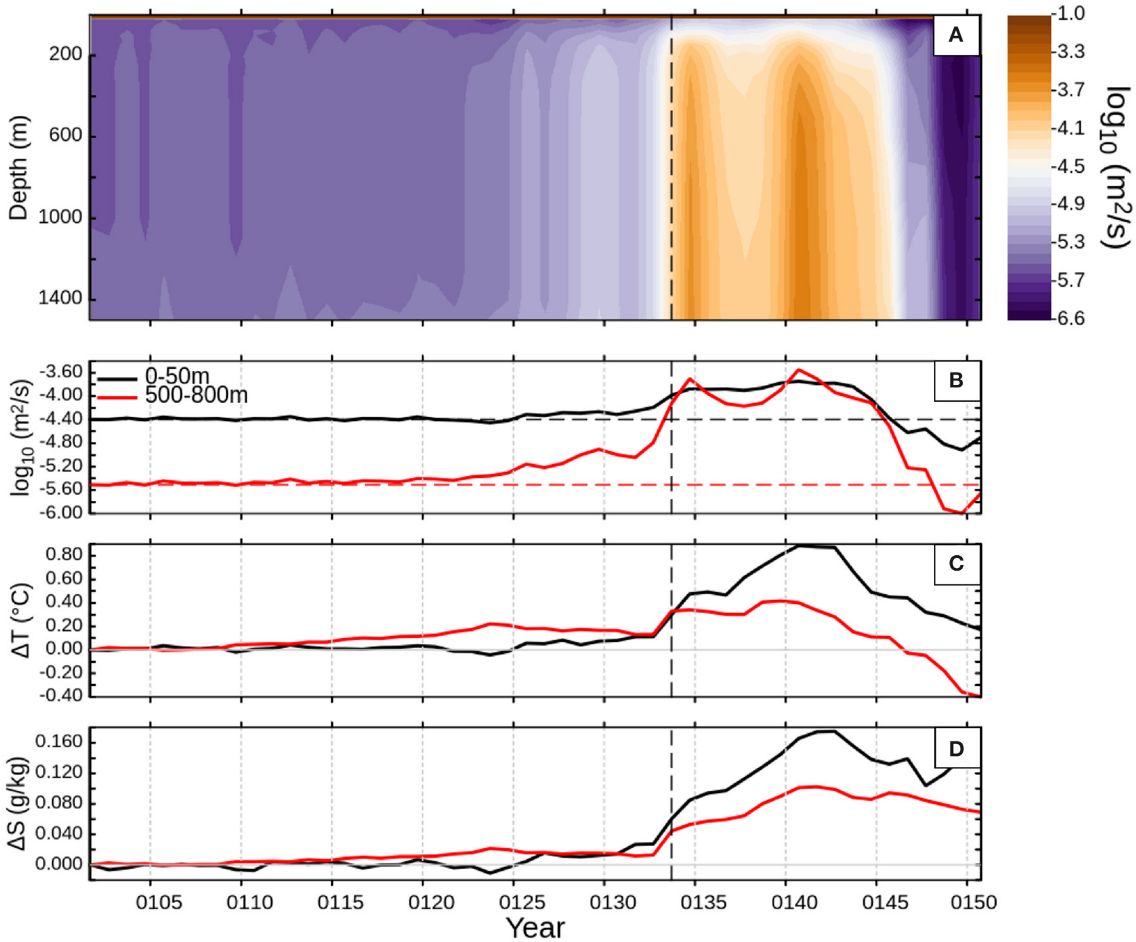

FIGURE 8 | (A) Hovmöller diagram of diapycnal diffusivity ($\log_{10}$ m^2 s^{-1}) for the upper 1500 m averaged over the polynya region. **(B)** Time series of September diapycnal diffusivity ($\log_{10}$ m^2 s^{-1}) with dashed lines corresponding to the mean pre-polynya values, **(C)** anomalies of potential temperature (°C), and **(D)** salinity (g/kg) for the upper 50 m (black) and at 500–800 m depth (red) in the polynya region. Anomalies are calculated with respect to a 10 year average from year 90 to 100, representing conditions at the end of the preconditioning phase. The opening of the first polynya is marked by the vertical dashed black line.

the stratification at these depths. The background stratification weakens in response, which again favors gravitational instabilities and leads to an erosion of the subsurface heat reservoir. This mechanism presents a positive feedback loop between diapycnal mixing and stratification; a weak stratification favors gravitational instabilities that enhance diapycnal mixing, which in turn weakens stratification further (Dufour et al., 2017).

3.5. Description of the Second Polynya Event

The second polynya forms further west in the Weddell Sea near 66°S, 45°W (**Figure 2B**; lower panel). However, the polynya formation mechanism is similar to the first event, e.g., preconditioning-mixing-convection, revealing a multi-centennial mode of polynya formation and decay (see also Goosse and Fichefet, 2001; Pedro et al., 2016).

After the first polynya closes in year 160, heat starts accumulating below the surface and the WDW reaches a maximum of 0.1°C after ∼100 years, similar to the build-up time preceding the first polynya. This corresponds to a mean warming rate of 0.013°C yr^{-1}, larger than the warming rate for the first polynya event (0.005°C yr^{-1}), but in agreement with observations (Robertson et al., 2002). This is due to the rapid resumption of CDW inflow resulting in an overshoot in $\theta_{\max}$ immediately after the first polynya closes (**Figure 4C**). Similar to the first event, the heat reservoir is maintained by thick and compact sea ice and a pronounced pycnocline which limits vertical exchanges between the WDW and the surface. A gradual deepening of the WDW core, through the preconditioning phase, also contributes to the build-up. We note that the pycnocline strength, indicated by the annual mean salinity difference between the mixed layer and the WDW (**Figure 9B**), weakens from ∼0.3 psu in year 100 to ∼0.2 psu in year 250. This makes the WDW more vulnerable to surface interactions and could potentially erode the heat reservoir and explain the lower $\theta_{\max}$.

Despite the lower WDW heat content it remains warm enough to melt sea ice and preventing ice formation in winter. Thus, when the second polynya appears at year 472

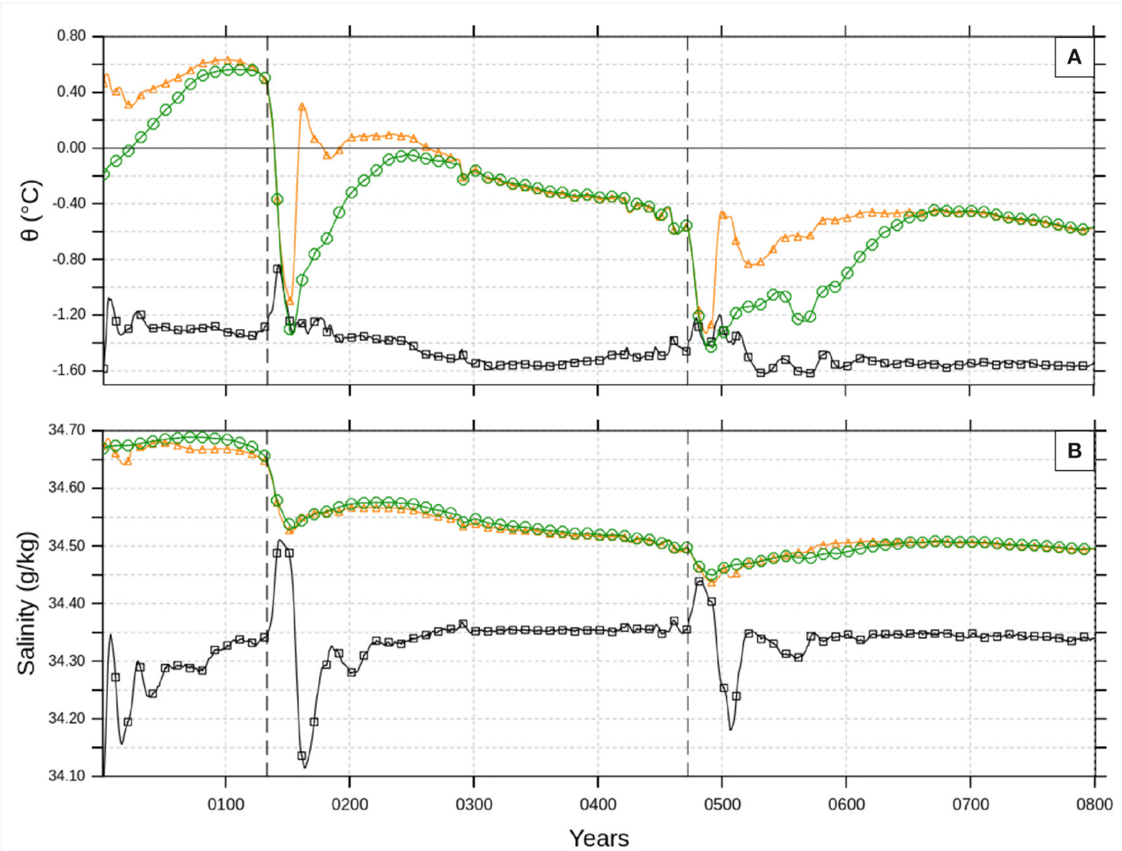

FIGURE 9 | Time series of annual mean **(A)** potential temperature (°C) and **(B)** salinity (g/kg) in the Weddell Sea Convective Region averaged between 31°W – 9°E and 60–70°S at the bottom of the winter mixed layer (black line with squares), the WDW core, e.g., θ_{max} (orange line with triangles) and below the WDW layer at 2,500 m depth (green line with circles). Vertical dashed lines mark the start of each polynya event.

(**Figure 2B**; lower middle panel) it triggers deep convection and the upwelling of WDW forces a rapid sea ice retreat (**Figure 2B**; lower right panel). A few years after the polynya opens, convection reaches its second peak and young surface water is visible at 4,000 m depth, while AOW near the surface gets older (**Figure 7**). Upwelling ceases in year 496 when the heat reservoir is depleted and sea ice recovers. The heat loss associated with the second polynya event is notably smaller than the first, due to the lower θ_{max} (**Figure 6B**). Interestingly, the polynya is similar in size and duration as the first polynya and therefore also triggers an increase in AABW of the same magnitude (**Supplementary Figure 4**). When the second polynya disappears, subsurface heat starts building up again, but only reaches a θ_{max} of -0.6°C (year 650). Consequently, no polynya forms in the remaining part of the simulation.

3.6. Cessation of Deep Convection

To illustrate how the simulated long-term changes in Weddell Sea water mass properties (WDW in particular) and stratification impact polynya occurrence we compare **Figure 9** with the vertical buoyancy profiles for each of the preconditioning phases shown in **Figure 10**. This may help in identifying key processes in polynya formation and understand why deep convection is absent in the later part of the simulation.

Figure 10 demonstrates that the stability of the water column (expressed in terms of buoyancy) is controlled mainly by salinity. Here, a fresh surface layer prevents cold water from sinking and the increase in salinity with depth is hindering warm and buoyant subsurface water in rising to the surface. The progressive cooling of the WDW (**Figure 9A**), therefore acts to stabilize the water column by reducing buoyancy below 1,000 m and weakens the convective potential. Meanwhile, the freshening of the WDW acts to increase buoyancy and weakens the background (haline) stratification at depth, as the salinity difference between the deep ocean (2,500 m) and the WDW diminishes (**Figure 9B**). Overall, the balance is in favor of increasing the total buoyancy of the WDW, making the water column less stable. As a consequence gravitational instabilities can be triggered more easily despite the lower θ_{max} value. Above the WDW layer, we find an increase in buoyancy associated with a freshening of the subsurface waters after the first polynya. This tends to stabilize the water column, working against the weaker stratification at depth (**Figure 10**),

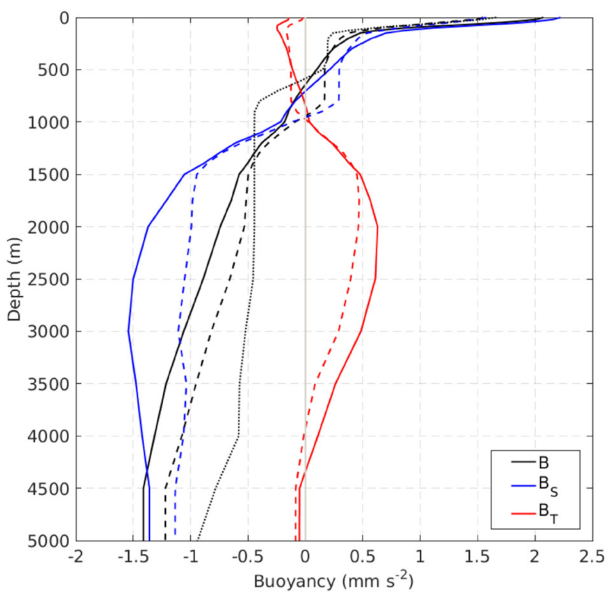

FIGURE 10 | Annual mean vertical buoyancy profiles (mm s^{-2}) averaged in the Weddell Sea Convective Region (outlined by green box in **Figure 1**) showing the total buoyancy (black), buoyancy due to the salinity term (blue), and buoyancy due to the temperature term (red) for years 100 (solid) and 250 (dashed). Contributions from temperature and salinity on the total buoyancy are calculated using a linearized equation of state. Here, the buoyancy is expressed as an anomaly with respect to the mean buoyancy profile at the end of each preconditioning phase. The dotted black line shows the total buoyancy at year 650 following the second and final polynya event.

and suppress convection for more than 200 years before the second polynya forms.

After the 2nd polynya event, the WDW cools and freshens even further (**Figure 9**). The water column below 1,000 m is weakly stratified with a salinity difference across the pycnocline of ∼0.15 psu, thus contributing to less heat and salt build-up at mid-depth. Meanwhile, the surface layer (upper 100 m) becomes progressively fresher, which acts to suppress deep convection (de Lavergne et al., 2014). Interestingly, a closer look at **Figure 9** shows that temperature and salinity increases above the WDW around year 700, indicative of enhanced vertical mixing. However, at this point the WDW is likely too fresh and too cold to erode the upper ocean stratification and trigger another polynya.

4. DISCUSSION

4.1. The Role of Mixing in Triggering Polynya Events

Ocean mixing plays a central role in the formation of polynyas simulated by the model, and is strongly linked to long-term changes in WDW properties and stratification. Climate model simulations by Kjellsson et al. (2015) and Timmermann and Beckmann (2004) demonstrate that the Weddell Sea stratification is highly sensitive to the parameterization of vertical mixing, thereby having a direct influence on polynya formation. Insufficient vertical mixing due to inadequate representation of buoyancy- and/or wind-induced mixing can lead to an accumulation of salt in the winter mixed layer that erodes the halocline, causing large and unphysical polynyas to form. Subsequently, it was suggested by Heuzé et al. (2015) that increasing vertical mixing in the upper ocean may reduce deep convection and prevent polynyas from forming. In contrast, we find that the polynya forms in response to enhanced vertical fluxes of heat and salt from below the mixed layer, associated with gravitational instabilities in the ocean interior (**Figure 8A**). This implies that increased diapycnal mixing is not related to changes in the surface forcing conditions (e.g., brine rejection, freshwater fluxes, or wind-stress), but rather forced by gradual changes in the WDW properties (**Figure 9**) thereby increasing the potential instability of the water column. A similar processes is also evident in coupled ocean-atmosphere models where the onset of deep convection is preceded by a gradual erosion of the subsurface heat reservoir over years or decades before the polynya opens (e.g., Martin et al., 2013; Zanowski et al., 2015; Vettoretti and Peltier, 2016; Dufour et al., 2017; Reintges et al., 2017; Zhang et al., 2019). What exactly triggers the convective instability is less clear, but Vettoretti and Peltier (2016) highlighted the role of double-diffusive mixing (not included in the NorESM-OC1.2) as a potential mechanism.

Whether this process plays a role in triggering the observed Weddell Polynya is more uncertain. Based on *in situ* observations for the 2017 Maud Rise polynya, Mojica et al. (2019) found that diapycnal and isopycnal mixing was important for initiating the polynya, with changes in the subsurface waters occurring before the polynya opened. Here, the enhanced bottom-up fluxes of heat and salt contributed to a higher level of instability in the subsurface waters, thus pointing to an oceanic preconditioning and supports that mixing processes plays an active role in polynya formation. This further highlights the importance of vertical mixing parameterizations to accurately simulate open-ocean deep convection in climate models (Timmermann and Beckmann, 2004; Heuzé et al., 2015; Kjellsson et al., 2015).

The vertical mixing also plays an important role for the build-up of the subsurface heat reservoir, which in turn affects polynya formation. Throughout the simulation we find a gradual decrease in WDW temperature and salinity (**Figure 9**) and a cessation of deep convection in the model. This could be linked to weaker stratification and enhanced levels of vertical mixing following each polynya event that prevents heat and salt in building up. This is supported by model simulations by Dufour et al. (2017), who found that heat accumulation relies on a stable stratification where vertical mixing is suppressed. They identified a particular role of representing dense water overflows and mesoscale eddies to simulate the stably stratified water column in the Weddell Sea. When stratification is too weak, arising from a poor representation of water masses, episodes of gravitational instability can trigger enhanced vertical mixing and lead to erosion of the WDW reservoir.

Another factor that may affect the subsurface heat reservoir is the deepening of the WDW layer. A shallower WDW core, is more vulnerable to upper ocean mixing, leading to increased ice-ocean heat fluxes that deplete the heat reservoir (e.g., Robertson et al., 2002). As evident in **Figure 4**, the simulated WDW is located around 1,000 m depth, and is therefore less influenced by surface forcing. We attribute the WDW deepening to a weak resupply of bottom water during the non-convective phases. When the polynya and deep convection is absent bottom water formation is reduced and we find a downward migration of the WDW core at a rate of 12 m yr^{-1} (**Figure 4**). This helps isolate the WDW and may also create favorable conditions for the formation of a polynya through cabelling and thermobaric effects (Akitomo, 1999; Mcphee, 2003; Wang et al., 2016). A similar response has been documented in other higher resolution models (e.g., Lee et al., 2002; Dufour et al., 2017) and represents an ongoing challenge influencing long-term climate model predictions. For example, Dufour et al. (2017) found an isopycnal displacement of 10 m yr^{-1} which was attributed to a lack of resupply of AABW and erosion of AABW properties. Alternatively, isopycnals models have been shown to be sensitive to the choice of the reference pressure, in particular in the Southern Ocean, (e.g., Assmann et al., 2010). Hence, the choice of 1,000 m as a reference level for our model density layers may cause stratification biases in the abyssal ocean and enhance consumption of bottom waters by spurious mixing processes.

4.2. What Controls the Duration and Frequency of Polynya Events?

We find two time scales associated with the frequency of open-ocean deep convection in NorESM-OC1.2. First, an advective-diffusive time scale associated with accumulation of subsurface heat, and second, a time scale set by the stratification and ocean mixing processes. The build-up of the heat reservoir occurs in about 100 years and is associated with advection of CDW into the Weddell Sea. Martin et al. (2013) show a similar time scale in the Kiel Climate Model and found that the centennial-scale recharge of the subsurface heat reservoir can drive an oscillatory mode of Southern Ocean deep convection, with the heat reservoir setting the time between polynya events. Meanwhile, observations suggest that already by the mid-1990s the WDW had recovered from the cooling induced by the 1970s convective event, which is significantly faster than the 100 years in our simulations (Robertson et al., 2002; Smedsrud, 2005). While observations remain too sparse to asses the full spatial extent of the WDW cooling by the Weddell Polynya, hydrographic profiles suggest that a significant amount of heat is left in the WDW. This lead to a faster recovery of the heat reservoir and the recurrence time for a Weddell Sea polynya is expected to be faster, than if the entire heat reservoir is depleted. Due to the large size of the simulated polynya, the total oceanic heat loss associated with convection is about 3.6×10^{22} J. This is an order of magnitude larger than the heat loss from 1974 to 1976 during the Weddell Polynya (Smedsrud, 2005) and is likely due to a combination of poor representation of sub-gridscale convection in coarse resolution models and the lack of atmospheric feedbacks. Despite the large simulated ocean-atmosphere heat fluxes associated with deep convection, the atmosphere remains cold and the strong surface buoyancy loss continues to drive upwelling of WDW until the heat reservoir is depleted. Including atmospheric feedbacks could help dampen convection by restratifying the water column, e.g., through changes in wind-driven upwelling, heat and moisture fluxes, or sea-ice transport into the polynya region (Martin et al., 2013; Gnanadesikan et al., 2020). Therefore, the 100-year time scale for the recharge process should be seen as an upper limit. Nevertheless, this result suggest a close link between the WDW heat content, polynya size and the frequency of polynya events: Large and long-lived polynyas leave the heat reservoir more depleted, the recovery time longer, and deep convection occurs less frequent. In contrast, for short-lived and smaller polynyas less heat is lost and the recharge time is shorter.

Changes in the local atmospheric conditions and ocean-atmosphere feedbacks are other process that could influence deep convection activity by affecting Weddell Sea stratification (e.g., Gordon et al., 2007; Gnanadesikan et al., 2020; Kaufman et al., 2020). For example, surface freshening by wind-driven advection of sea ice, or precipitation anomalies, acts to stabilize the water column and suppress polynya formation (Comiso and Gordon, 1987; de Lavergne et al., 2014). On the other hand, recent analysis of the 2016 and 2017 polynya event, indicate that variations in Southern Hemisphere winds were important for triggering the polynya, by creating ice divergence and wind-driven upwelling of WDW (e.g., Cheon et al., 2015; Campbell et al., 2019; Francis et al., 2019). Such variations may be linked to long-term changes in large-scale atmospheric patterns like the Southern Annular Mode (Gordon et al., 2007; Jena et al., 2019). However, since the atmospheric forcing is fixed in our simulation, there is no external variability in the surface forcing conditions (i.e., winds, precipitation) that could impact polynya activity. Instead, changes in stratification are driven by internal ocean processes (see also Dufour et al., 2017) and thus presents a purely oceanic time scale for polynya formation. This absence of stochastic forcing by the atmosphere likely leads to polynyas forming less frequently. Therefore, we speculate that the oceanic processes described here are dominating polynya activity on centennial time scales (through advective-diffusive processes) whereas atmospheric variability is important over shorter inter-annual to decadal time scales.

4.3. Concluding Remarks

Using a stand-alone ocean-sea ice model (NorESM-OC1.2), we show that internal ocean processes can trigger the formation of open-ocean polynyas and drive deep convection in the Weddell Sea, without changes in the atmospheric forcing. The proposed mechanism relies on the build-up of WDW as an essential preconditioning for polynya formation (Cheon et al., 2015; Dufour et al., 2017), leading to gravitational instabilities in the ocean interior and enhanced vertical mixing.

The polynya formation mechanism can be separated into three phases: (1) a preconditioning phase characterized by accumulation of WDW at mid-depths sustained by a strong stratification that suppresses vertical mixing; (2) a mixing phase associated with enhanced upward fluxes of heat and salt leading to a weakening of the upper-ocean stability and a small polynya opening; and (3) a convective phase, where deep convection drives upwelling of WDW that maintains the polynya. After a couple of decades, the subsurface heat reservoir is depleted and the system returns to phase (1).

Finally, we show that deep convection causes significant changes in Weddell Sea deep and bottom water properties which eventually leads to a shut-down of deep convection in the model. This provides a perspective on the long-term evolution of Southern Ocean polynya activity. We find that the polynya cannot form when the WDW becomes too cold ($\theta_{max} < -0.6°C$). In comparison, Cheon et al. (2014) found that open-ocean polynyas can form even for very low θ_{max} values close to the freezing point (−1.4°C), implying that additional factors (such as the salinity of WDW) might be important for initiating the polynya. Here we show that the destratification by subsurface ocean mixing is prevented when the WDW is freshened, thereby suppressing ocean-ice heat fluxes and sea ice melt. Therefore, we propose that the combination of a colder and fresher WDW may lead to a cessation of deep convection. As opposed to the changes simulated here, the past decades have shown a warming of the Southern Ocean deep waters (e.g., Purkey and Johnson, 2010; Fahrbach et al., 2011), which could help destabilize the water column and favor deep convection. At the same time global climate models predict convection to decrease in the twenty-first century due to a continued freshening of Southern Ocean surface waters under warming conditions (de Lavergne et al., 2014). Given that the warming trend of WDW is enough to overcome the enhanced surface stratification, the likelihood of large and long-lived Weddell polynyas occurring in the future could therefore increase.

DATA AVAILABILITY STATEMENT

The raw data supporting the conclusions of this article will be made available by the authors, without undue reservation.

AUTHOR CONTRIBUTIONS

The analysis, plotting and preparation of the manuscript was done by JR. LS and KN also aided with the interpretation of the model simulation. All authors contributed to the writing and editing of the final manuscript.

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SUPPLEMENTARY MATERIAL

The Supplementary Material for this article can be found online at: https://www.frontiersin.org/articles/10.3389/fclim.2021.718016/full#supplementary-material

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frontiers in Climate

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The Role of Blue Carbon in Climate Change Mitigation and Carbon Stock Conservation

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The potential for Blue Carbon ecosystems to combat climate change and provide co-benefits was discussed in the recent and influential Intergovernmental Panel on Climate Change Special Report on the Ocean and Cryosphere in a Changing Climate. In terms of Blue Carbon, the report mainly focused on coastal wetlands and did not address the socio-economic considerations of using natural ocean systems to reduce the risks of climate disruption. In this paper, we discuss Blue Carbon resources in coastal, open-ocean and deep-sea ecosystems and highlight the benefits of measures such as restoration and creation as well as conservation and protection in helping to unleash their potential for mitigating climate change risks. We also highlight the challenges—such as valuation and governance—to marshaling their mitigation role and discuss the need for policy action for natural capital market development, and for global coordination. Efforts to identify and resolve these challenges could both maintain and harness the potential for these natural ocean systems to store carbon and help fight climate change. Conserving, protecting, and restoring Blue Carbon ecosystems should become an integral part of mitigation and carbon stock conservation plans at the local, national and global levels.

Keywords: ecosystem services, mitigation, carbon services valuation, governance, environmental economics

INTRODUCTION

The ongoing global COVID-19 pandemic and its global human and economic repercussions brings to the fore the recognition that the sustainability of our economic systems very much depends on the sustainability of ecosystems and biodiversity. The current climate change crisis is threatening economies as it is accelerating losses of marine biodiversity and habitats (Bindoff et al., 2019). There is an increased awareness of the severe impact of damage to the natural world on social and economic well-being, and a growing urgency in calls to make (or demand) changes that will put societies on a more sustainable path. To date, conservation appeals have not attracted investment in natural capital at the level needed, nor have appeals to focus on mitigating the effects of the climate crisis. There is a dire need for a change in societal mindsets toward those that recognize nature as invaluable to our economic well-being.

The pandemic has shaken a number of assumptions common to modern societies regarding the relationship between people and the natural world—an extractive view of nature—exemplifying the high cost and great vulnerability inherent in this outdated mentality. International lockdowns revealed the positive effects of reduced anthropogenic disturbance on natural ecosystems. Governments and industries have been able to adapt rapidly and radically to mitigate the worldwide pandemic risk to societies. This has laid bare the old arguments that rapid adaptation is not possible at the scale needed to reduce the threats of climate change.

Marine ecosystems require far greater attention than received thus far as a means of securing humanity's future health and well-being (Laffoley, 2020; Laffoley et al., 2020). Those marine ecosystems that contribute to climate change mitigation by sequestering excess carbon from the atmosphere are known as Blue Carbon ecosystems. The Intergovernmental Panel on Climate Change (IPCC) defines Blue Carbon as "All biologically-driven carbon fluxes and storage in marine systems that are amenable to management." The focus has been on rooted vegetation in the coastal zone, such as tidal marshes, mangroves and seagrasses. These ecosystems have high carbon burial rates on a per unit area basis and accumulate carbon in their soils and sediments. They provide many non-climatic benefits and can contribute to ecosystem-based adaptation to climate change. If degraded or lost, coastal blue carbon ecosystems are likely to release most of their carbon back to the atmosphere. There is current debate regarding the application of the blue carbon concept to other coastal and non-coastal processes and ecosystems, including the open ocean." (Weyer et al., 2019).

Many natural processes and ecosystem components contribute to carbon sequestration and burial; when these are disrupted additional carbon previously stored can be released into the ocean or atmosphere. We refer to protection of these processes and stores as *carbon stock conservation,* an action that is transitional between mitigation and adaptation. There is a gradient in anthropogenic influence on natural ocean carbon stocks and their value ranging from destruction and degradation (inducing carbon loss) to restoration and creation (enhancing carbon) with maintenance and protection of existing carbon storage as neutral (**Figure 1**). Despite their importance, these blue carbon concepts are not yet uniformly incorporated into climate strategies on local, national and global scales. The reasons for this include variability in ecosystem impacts, uncertain carbon fluxes and trajectories, valuation methods and governance strategies. Resolving these challenges would go a long way in enhancing the efforts to include these natural solutions to climate change into both mitigation and carbon stock conservation strategies, where they can be beneficial to a country's sustainable development.

The Special Report on the Ocean and Cryosphere in a Changing Climate (SROCC) (IPCC, 2019) provides policy makers and other parties with a holistic perspective on the current state of the ocean in the face of increasing problematic impacts of climate change. The report identifies knowledge gaps and uncertainties which limit the design and implementation of mitigation strategies in the UNFCCC policy framework. It introduces Blue Carbon-driven ecological and economic climate change mitigation measures and identifies two management approaches in Chapter 5 (Bindoff et al., 2019) (i) Actions to maintain the integrity of natural carbon stores, decreasing their potential release of greenhouse gasses and (ii) Actions that enhance the long-term removal of greenhouse gasses from the atmosphere by marine systems. SROCC briefly discusses ocean processes such as the biological and microbial pumps that transform and transport carbon, and reviews approaches to ocean management for climate presented in Gattuso et al. (2018). SROCC highlights the release of carbon due to disturbance to coastal vegetated habitats, the need for habitat protection, and refers to the mitigation potential of habitat protection. It also discusses restoration, creation and maximizing carbon uptake and storage of coastal ecosystems and highlights complexities in defining effectiveness of these mitigation actions, concluding that the overall potential to mitigate climate change is low (0.04 to 0.05 Gt/y) or < 0.5 % of 10 Gt/y of anthropogenic emissions from all sources.

This paper expands the discussion of the role of Blue Carbon in climate change mitigation strategies by focusing on the importance of conserving existing marine pathways of carbon fixation, transport, burial and sequestration, and highlighting the challenges associated with the measurement, valuation, management, and governance of carbon in coastal, open ocean, and deep-sea ecosystems. Section Ecological and Economic Aspects of Blue Carbon identifies the carbon sequestration services provided by these ecosystems. Section Management of Blue Carbon for Sustainable Economies highlights how valuing ecosystem services (ES) can support release of financial resources from the private sector to help develop markets around the protection and regeneration of blue carbon, nature-based solutions, and Section Conclusion concludes with policy recommendations.

ECOLOGICAL AND ECONOMIC ASPECTS OF BLUE CARBON

As Blue Carbon becomes a center of focus for research and policymakers, improvement is needed in our ability to measure rates and permanence of carbon sequestration (Macreadie et al., 2017). Anthropogenic threats to the integrity of Blue Carbon stocks, defined here as carbon sequestered in the ocean, were not fully addressed by the SROCC. These threats, however, must be identified and weighed against the advantages of protecting these carbon draw-down systems so as to develop a sustainable economy. So far, governments, local communities and private sectors have worked with little cohesion, but once Blue Carbon ecosystems are clearly defined and their social and economic vulnerabilities to anthropogenic change are pinpointed in IPCC reports, they can then be harnessed to work in tandem toward sustainable goals to mitigate the socio-economic costs and ecological impacts of a changing climate. The following sections identify the potential for incorporating Blue Carbon in mitigation and carbon conservation strategies from both an

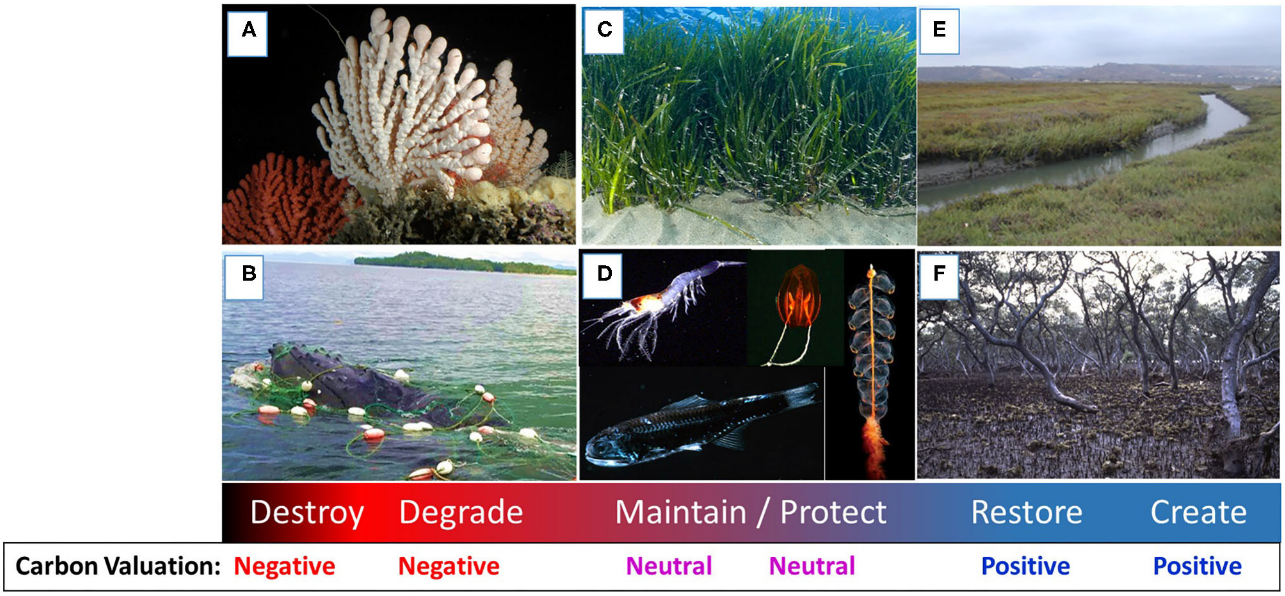

FIGURE 1 | Examples of ecosystems and processes along a gradient of actions that can have negative, neutral or positive effects on carbon storage. **(A)** Demersal fishing gear has destroyed many ancient deep-sea coral reefs in the past 30 years, and none have been restored (photo credit JAGO-Team/IfM-GEOMAR). **(B)** Whale populations have been decimated by hunting, ship strikes or entanglement in plastic fishing nets, lowering their ability to transfer carbon from the surface ocean to the deep-sea. **(C)** Seagrass beds can store large amounts of carbon but have been in decline worldwide, requiring protection for those that remain (credit C Vasapollo). **(D)** Pelagic animals such as mesopelagic fish draw carbon down from the shallows to great depths, their populations need to be maintained to protect this ecosystem service. Coastal developments have removed most saltmarshes **(E)** and mangrove forests **(F)** worldwide, requiring efforts to restore or recreate these important natural carbon stores.

ecological and economic standpoint to help design sustainable management frameworks.

Blue Carbon Ecosystems and Their Functions

Coastal Ecosystems

The term Blue Carbon was coined a decade ago to describe the disproportionately large contribution of coastal vegetated ecosystems to global carbon sequestration and the need to protect these resources (Macreadie et al., 2017). Mangrove forests, saltmarshes and in some cases, seagrasses, build up large stocks of organic carbon in the coastal zone as they grow in depositional soils. Carbon storage in the soils of marine angiosperm (higher plant) habitats can be up to 1,000 tC ha^{-1}, much higher than most terrestrial ecosystems (IPCC, 2019). Rising atmospheric CO_2 levels are causing ocean acidification, but the increased amount of CO_2 in seawater can stimulate photosynthesis helping to remove carbon from the seawater (Wada et al., 2021). Hence, another benefit of conserving submerged Blue Carbon habitats is that they may ameliorate ocean acidification locally (Su et al., 2020).

Increasing population densities and urbanization of coastal areas has damaged vegetated coastal habitats worldwide due to the impacts of fisheries, aquaculture, pollution and sedimentation (Gullström et al., 2021). Around 62% of mangroves worldwide were destroyed between 2000 and 2016 (Goldberg et al., 2020), there has been about a 90% loss of salt marsh ecosystems (Gedan and Silliman, 2009) and seagrass carbon stocks are declining in various regions of the world (Waycott et al., 2009). When these blue carbon stocks are damaged, they may switch from sinks to sources of CO_2 and of methane (CH_4) to the atmosphere, a much more potent greenhouse gas (Hiraishi et al., 2014; Macreadie et al., 2017; Vanderklift et al., 2019). Even without damage, some coastal wetland systems may emit significant methane and nitrous oxide, thus additional research is needed on how to manage coastal ecosystems for maximum carbon sequestration benefits (Rosentreter et al., 2021). Ocean warming affects the ability of marine systems to remove CO_2 from the atmosphere because warmer waters absorb less CO_2 and because it is a stressor for cool water vegetated marine habitats. For example, kelp forests in the warmest part of their NE Atlantic distribution store around 70% less carbon and release 50% less carbon than populations in the cooler parts of their distribution (Pessarrodona et al., 2018). Marine forests of kelp and fucoid seaweeds are being lost globally at their low latitude boundaries due to marine heatwaves and the gradual warming of surface seawater temperatures combined with human- induced stressors (Merzouk and Johnson, 2011; Kletou et al., 2018; Bernal-Ibáñez et al., 2021).

The SROCC report (IPCC, 2019) made clear that the maximum mitigation benefits of mangrove, seagrass and saltmarsh restoration is unlikely to reach more than 2% of current total CO_2 emissions, but improved protection and management of these critical habitats would have multiple benefits. These include providing storm protection, improving water quality, benefiting biodiversity and fisheries as well as reducing carbon emissions from these ecosystems (Windham-Myers et al., 2018). The SROCC report did not fully address the importance of Blue Carbon sources and sinks beyond marine angiosperm habitats. For example, these habitats export carbon into the open ocean, so measurements of the rate of carbon build up in angiosperm habitats may underestimate the role they play in carbon sequestration (Santos et al., 2021).

Beyond Blue Carbon hotspot habitats, carbon is also stored in marine animals, so fish taken out of the sea contribute to Blue Carbon release (Mariani et al., 2020). Bottom trawled fishing gear damages ancient Blue Carbon stores such as rhodolith/maerl beds in seafloor sediments (Riosmena-Rodriguez et al., 2017) and is estimated to release a gigaton of CO_2 from seabed sediments each year, equivalent to the entire aviation industry's annual emissions to the atmosphere (Sala et al., 2021). In sandy sediments, bottom fishing kills organisms that regulate carbon cycling on the seafloor (Hale et al., 2017), whereas in mud habitats it also releases carbon stored in the sediment itself (Sciberras et al., 2016). Bottom trawling also decreases the flux of organic carbon from shallow coastal waters to the deep sea by over 60% in North Western Mediterranean waters (Pusceddu et al., 2014; Laffoley, 2020).

Open Ocean Ecosystems

The ocean has dissolved carbon stocks that are at least an order of magnitude greater than those in global terrestrial soils. Most of this dissolved carbon is bicarbonate with an ocean residence time of around 100,000 years (Millero, 2007). Oceanic dissolved organic carbon is nearly equal to atmospheric CO_2 and about 200 times the amount of carbon found in living marine biomass (Worden et al., 2015). Marine phytoplankton are responsible for ∼50 % of global primary production (∼50 Gt C/year), these photosynthetic algae and bacteria fix dissolved inorganic carbon which is then mainly consumed and stored in the biomass of other organisms. The amount of carbon that is fixed by phytoplankton and then sequestered varies regionally and temporally, depending on surface water productivity, grazing/microbial degradation, and physical processes such as turbulence (Barnes et al., 2020; Briggs et al., 2020).

Most of the carbon fixed by phytoplankton is grazed by zooplankton although viral attack can also release their organic material which is then broken down by other microbes (Breitbart et al., 2018) generating particulate and dissolved organic matter, some of which will end up in deep water masses. The microbial breakdown of organic carbon produced by phytoplankton sequesters nearly 0.2 Pg C/year into the deep sea (Legendre et al., 2015) and around 30% of the organic particles from the sunlit ocean are exported below the mesopelagic zone (Briggs et al., 2020). The rates of sinking and degradation of carbon from phytoplankton are affected by cell size, morphology and chemical composition (Bach et al., 2019; Richardson, 2019). Fecal pellets, exoskeletons, dead animals and the vertical migrations of open ocean animals also transport the carbon from phytoplankton into the deep sea (Barnes and Tarling, 2017; Boyd et al., 2019).

Ocean warming and acidification are projected to slow sinking of particulate organic carbon to the deep seafloor by about 10–15% by the end of this century under a high CO_2 emission scenario due to a projected decrease in primary production (Bindoff et al., 2019) and a community shift toward phytoplankton with smaller cells (Flombaum et al., 2020). There is uncertainty in this projection and so research into open ocean carbon sequestration rates is needed to locate and understand hotspots of open ocean carbon sinks (Gattuso et al., 2018; Buesseler et al., 2020; Jin et al., 2020; Martinetto et al., 2020).

Over the past 6 years macroalgae in the genus *Sargassum* has bloomed in open ocean areas of the North Atlantic and Caribbean. This has ended up in massive quantities on beaches with negative impacts on tourism, human health and coastal ecology (Van Tussenbroek et al., 2017; Resiere et al., 2018; Chávez et al., 2020; Gouvêa et al., 2020; Landrigan et al., 2020). Dealing with this has been expensive, Mexico, for example, declared a state of emergency and spent $17 million in 2018 to remove 500,000 tons of *Sargassum* from its Caribbean beaches (Landrigan et al., 2020). Others have suggested *Sargassum* as a source of food or food additives (Amador-Castro et al., 2021; Choudhary et al., 2021), clean energy (Amador-Castro et al., 2021) and fertilizer (Thompson et al., 2020), or a local antidote to ocean acidification in bivalve culture (Han et al., 2020).

Sargassum (floating and attached to the seafloor) occurs in such large quantities globally (13.1 PgC) and covers such vast areas (445.54×10^4 km^2) (Gouvêa et al., 2020), that its carbon sequestration capacity has drawn attention globally (Krause-Jensen and Duarte, 2016; Hu et al., 2021; Santos et al., 2021) and regionally (e.g., Korea-Sondak and Chung, 2015), but with conflicting perspectives. Hu et al. (2021) suggest that the C fixed and stored in *Sargassum* in the North Atlantic (6 M tons /mo) represents <0.2% of phytoplankton storage, and is not globally significant, although it may have local importance. About 10% of the surface production of Atlantic *Sargassum* reaches the deep seafloor as particulate organic matter, but massive episodic inputs can occur during storms (Krause-Jensen and Duarte, 2016). Wild *Sargassum* growth and sinking has been proposed as a natural analog of basin-scale seaweed farming (afforestation), one method of enhanced carbon dioxide removal under consideration, because it involves scales that are orders of magnitude larger than feasible pilot studies. However, Bach et al. (2021) suggest that nutrient reallocation (from phytoplankton to *Sargassum*) and calcification by encrusting marine life reduce the net carbon dioxide removal efficiency of *Sargassum* by 20–100% and that increased ocean albedo of *Sargassum* could have a larger effect on climate radiative forcing than *Sargassum* removal.

Deep-Sea Ecosystems

Dissolved organic carbon is ∼70% of the total organic carbon in the ocean, and most of this is found at depths >1, 000 m where this carbon remains out of contact with the atmosphere for thousands of years (Hansell et al., 2009). Mesopelagic zooplankton and fish typically migrate large distances each day to optimize their own feeding and to avoid predators and this plays a major role in transporting carbon down from surface waters

(Davison et al., 2013; Steinberg and Landry, 2017; Kiko and Hauss, 2019). Deeper water fish then transfer carbon into long-term storage below 1,000 m depth (Trueman et al., 2014) where most of this carbon remains sequestered from the atmosphere for thousands of years. This has led to suggestions to grow and deposit macroalgae in deep water (Duarte et al., 2017; Queirós et al., 2019) as well as rebuild stocks of marine vertebrates (Martin et al., 2016; Smale et al., 2018). There is also interest in conserving processes that transport carbon into deep water e.g., vertically migrating mesopelagic krill or calcifiers that provide ballast (Howard et al., 2017b). As in shallow waters, carbon can be sequestered via passive burial in sediment or by active bioturbation (Atwood et al., 2020). The carbon stocks in the top 1 m of seafloor sediments (3,117 Pg) are more than twice that of terrestrial soils (Atwood et al., 2020).

Rising bottom temperatures or shifting of warm currents could increase the release of carbon stored in buried methane hydrates on continental margins (Ruppel and Kessler, 2017). Chemosynthetic ecosystems, and methane seeps, in particular, sequester carbon compounds emitted from within the Earth's crust (Thurber et al., 2014). Microbes convert methane into carbonate on the seafloor and also make this carbon source available to animals as symbionts. Ocean warming may dissociate more methane, and if this occurs, ensuring it never reaches the atmosphere takes on added importance. This sequestration service has never been subject to monitory valuation or even justified seep conservation, but the potential for novel CO_2 and CH_4 removal mechanisms of potential climate remediation value has been recognized (e.g., Diaz-Torres et al., 2015; Mahon et al., 2015).

Just as trawling releases carbon from shelf and slope marine sediments, deep-seabed mining (which has yet to occur) could release carbon sequestered in sediments, although some of this material would still be in waters >1,000 m deep and not released to the atmosphere. In the case of mining of polymetallic nodules, the carbon content of these sediments is very low as they occur below oligotrophic waters and so the risk of releasing carbon to the atmosphere is likely to be small (Levin et al., 2020). This may not be true on organic-rich continental margins (e.g., areas targeted for phosphate mining), where remineralization could stimulate more phytoplankton production (Atwood et al., 2020).

The ocean plays a critical role in global climate regulation through uptake and storage of heat and carbon dioxide. However, this regulating service causes warming, acidification and deoxygenation and leads to decreased food availability at the seafloor (Levin and Le Bris, 2015; Sweetman et al., 2017). These changes are likely to affect the productivity, biodiversity and distributions of deep-sea fauna, thereby compromising key ES (Mora et al., 2013; Sweetman et al., 2017; Morato et al., 2020). Limited information on deep-sea species thresholds, tolerances and tipping points for various climate drivers means that predictions of risk, vulnerability and responses are difficult to make, and confidence is low. To date, the most knowledge involves projected changes in biomass in response to declining surface productivity and POC flux (Yool et al., 2017; Jones et al., 2018; Bindoff et al., 2019).

Policy Implications

Beyond Carbon Benefits

Coastal ecosystems are by definition a highly active interface between human and natural infrastructures which is exposed to a number of potentially threatening human activities. These activities include aquaculture, fisheries, coastal tourism, coastline development/habitat degradation and waste-water discharges which accompany growing human density on coastal shelves. Such impacts have recently been shown to have long-term deleterious effects such as the decrease of tidal marsh carbon sediment stocks due to human reclamation (Ewers-Lewis et al., 2019). While the active integration of Blue Carbon ecosystems into sustainable policy frameworks supports natural CO_2 entrapment, it also could allow for indirect monitoring of these anthropogenic disturbances.

Coastal Blue Carbon ecosystems also present a key advantage by supplying multiple ecosystem services (ES) in addition to carbon sequestration, upon which climate and population security might rely (Windham-Myers et al., 2018; Duarte et al., 2020). Key ES provided by coastal carbon ecosystems include protection of coastal habitats which serve as feeding and nursery grounds for fish and shellfish, protection of coastal infrastructure (for transportation, communication, dwelling, energy etc.) from storm surge and flooding as well as provision of water filtration. These ES are crucial for vulnerable communities that live near the shoreline or rely heavily on resources from these ecosystems[1]. Connectivity between ecosystem services is a key challenge to policymakers along with site- and species-specific requirements. Although providing <1% of global GHG, small islands are highly exposed to problematic impacts of climate change: changing temperatures, OA, weather disturbances are some of the many (Wilson and Forsyth, 2018). Viable policy frameworks which incorporate Blue Carbon ecosystems can mitigate climate change conflicts by sustainably using carbon ecosystems to support vulnerable communities, a method known as ecosystem-based adaptation (EbA) and a type of nature-based solution (NbS). Nature-based Solutions contribute to climate change mitigation on 3 fronts: reduction of GHG emissions, carbon capture and storage and socio-economic benefits of mitigation strategies (Raghav et al., 2020). They can physically shield communities effectively and can be a key resource for survival: wetlands not only mitigate the impacts of floods and storms but can also provide water which is naturally stored for communities in need. Attempts at replacing this protective service with man-made infrastructures have met with little success, high costs, and continuous ecosystem depletion.

Assessing and Quantifying

With the array of carbon ecosystems comes an array of processes by which carbon is sequestered from the medium from which it is extracted (let it be air or water). This service is too often poorly understood as there is both a flow of carbon passing through this natural machinery (through the

[1]https://www.worldbank.org/content/dam/Worldbank/document/EAP/Pacific%20Islands/climate-change-pacific.pdf

process of sequestration) as well as carbon stocks in relevant ecosystems (Keith et al., 2021). The real absolute quantity of carbon trapped in ecosystems can thus be poorly accounted for if these two phases are not taken into consideration. Often younger ecosystems are given priority while key older ones are destroyed along with their carbon storage. In order to develop effective policies for climate change mitigation, carbon stock assessments ought to be considered (Gullström et al., 2021).

Carbon flow through open marine ecosystems is a cycle in which living components have dynamic movements. Thus, carbon accounting must also avoid underestimating the transfer of allochthonous carbon, which has by definition traveled from its source habitat. Policy has commonly focused on carbon ecosystems in isolation to facilitate management without taking this transfer from source habitats into account. A contribution to sustaining the sequestration of marine carbon is through the conservation of marine vertebrates who support this cycle by transferring carbon from surface to deeper waters (Smale et al., 2018).

Monitoring and Protecting

Once identified, carbon-sequestering ecosystems must be a point of focus for both mitigation and conservation efforts, whether on local, national or global scales. For example, with shorter sea-ice durations and higher surface temperatures, entire areas of the Antarctic seafloor are offering a new alleyway for carbon sequestration, currently with very limited anthropogenic disturbance, causing highly productive benthic communities (Fillinger et al., 2013). However, human exploitation must be kept to a minimum through international policy efforts that should stand firmly in the face of commercial industry interests which would seek to exploit this newly freed and somewhat pristine part of the Southern Ocean for living and non-living resources (Bax et al., 2020). Gogarty et al. (2020) propose "non-market" approaches to achieve some Paris Agreement goals in managing this new area. Market solutions that aim to highlight the value of regenerative natural systems can also be applied (Chami et al., 2019). Policies based on a similar thinking ought to also be applied to high-seas and intertidal areas which are too often cast aside in MPA design.

While key ecosystems must be identified and protected, artificial mechanisms can also be used to support their impact on climate change mitigation. Negative Emissions Technologies (NET) might offer short-term mitigation, but their climate and environmental costs and benefits are yet to be ascertained for projects of sufficient scale to mitigate climate change. Only ocean fertilization has been reviewed so far and ascertained as ineffective in the short term (Williamson et al., 2012). Others include enhanced weathering, reforestation, bioenergy production with carbon capture and storage (BECCS), carbon fixing in soils as well as direct air carbon capture and sequestration (DACCS) which has the advantage of presenting less adverse impacts (Gambhir and Tavoni, 2019). However, these systems can be water and energy intensive and even require vast dedicated land over large timescales (with negative biodiversity effects), thus they must be approached with caution[2]. Similarly, any negative impact of these systems on key carbon ecosystems, a process known as "leakage," must be monitored and avoided (Ullman et al., 2013). NET are not alternatives for emission reduction or nature-based carbon conservation: they must imperatively be paired with strategies that protect carbon-sequestering ecosystems[2].

Preserving and Restoring Carbon Sequestering Ecosystems

According to Pendleton et al. (2012), emissions from the degradation of these ecosystems are equivalent to 3–19% of deforestation worldwide and result in economic damages of around 6 to 42 billion US$ annually. The degradation of coastal ecosystems each year releases between 0.15 and 1.02 Pg (one billion tons) of CO_2 into the atmosphere. Policy makers can shield key carbon cycles which are disturbed by anthropogenic activity through conservation frameworks. Although Marine Protected Areas (MPA) are growing at about 8% per year (Worm, 2017), this protective system remains far too limited with below 3% of the global ocean being effectively protected. Ocean protection could yield several benefits alongside securing carbon stocks at risk from bottom trawling, such as supporting fisheries' yield and protecting biodiversity (Sala et al., 2021).

Once disturbed, some functions of these ecosystems can still be restored and maintained. As mentioned above, ES can ensure climate security for local communities and restoration of carbon ecosystems has the potential to optimize these benefits. For nations that are small greenhouse gas emitters, addition of Blue Carbon habitat may offset their emissions as contributions to the Paris Agreement, but this cannot substantially address the global greenhouse gas problem. However, once policy frameworks incorporate the challenges of restoration program (mainly continuous funding, site- and species-specific requirements and extensive time frames (Wilson and Forsyth, 2018), they can have vast positive impacts on both habitats and populations. The valuation of protecting and restoring coastal ecosystems not only creates financial incentives while attracting investor interest but can also entail long-term investments into community development. As Cziesielski et al. (2021) explain regarding Red Sea Blue Carbon ecosystems, policy design must require the involvement of different fields of expertise (financial, technological, sociological and ecological) in order to ensure optimal long-term economic benefits, social support and ecological conservation. This is the founding principle of a sustainable Blue Economy[4]. In the current world, the only way to design such policy actions is to understand the economic powers behind the scenes: 47% of factors constraining policy makers have recently been shown to be financial in nature (Beeston et al., 2020).

[2]https://easac.eu/fileadmin/PDF_s/reports_statements/Negative_Carbon/EASAC_Report_on_Negative_Emission_Technologies.pdf
[3]https://mpatlas.org
[4]See https://www.worldbank.org/en/programs/problue.

MANAGEMENT OF BLUE CARBON FOR SUSTAINABLE ECONOMIES

Valuing Carbon Services

Coastal Carbon

Coastal marine ecosystems could provide as much as two-thirds of the ecosystem services that make-up our planet's natural capital (Cantral et al., 2012). Nevertheless, these services have been neglected through inadequate management, misled governance and gaps in social and scientific knowledge along with lack of local knowledge. A study reported by Quevedo et al. (2021) documented that coastal ecosystem services are more likely to be acknowledged by a society which benefits directly from them; in other words, clearer pathways between ecosystem services and societies are key for populations to truly value an ecosystem as a whole. This explains partly why Pacific Island nations have been at the forefront of advancing ocean issues in climate policy. For example, Chami et al. (2020b), provide a value for the carbon sequestration services of the existing saltmarshes in England. They use estimates of the carbon sequestration of the remaining saltmarshes as well as their flood control services to derive a value of $4.7 billion dollars. Sun and Carson (2020) quantified the value of storm protection from Gulf of Mexico wetlands at an average of $1.8 million dollars per year per km^2 (carbon sequestration services were not a part of this valuation). Recent discoveries in Red Sea mangrove forests suggest an underestimation of carbon sequestering potential due to the unaccounted-for positive impact of ocean acidification on the ocean's capacity to dissolve CO_2 (Saderne et al., 2020), a process likely applicable for a number of key areas around the globe.

The economic value of ecosystem services is either determined from quantifiable resources directly derived from the ecosystem, a service which is vulnerable to disturbance or changes over time (Fisher et al., 2008), or from the cost of restoring these coastal habitats were destruction to occur (Garrod and Willis, 1999). Payments for ES aim to incorporate the socio-economic levels involved in ES valuation but are sensitive to the highly changeable market of carbon pricing as well as issues of accountability and governance (Börner et al., 2017). ES valuation has been criticized as it might overlook the complexity of the connectivity between services and non-pecuniary services of ecosystems regarding their aesthetics and importance in local culture (Kosoy and Corbera, 2010). This might shift the focus of authorities from long-term social and environmental benefits to mainly financial returns. This perception of ES value can also change with different stakeholders and from regional to global scale (Brown and Adger, 1994).

Open Ocean Carbon

From a welfare point of view, marine systems provide a regulatory carbon service with global impact. The Exclusive Economic Zones (EEZs) in the Mediterranean basin have been used to estimate the wider economic impact of marine carbon sequestration in this area (Canu et al., 2015). The results indicate that the Mediterranean Sea is a key global sink of CO_2 with an estimated overall flux of CO_2 of 17.8 million-ton CO_2/year. The EEZs of Algeria, Greece, Italy and Spain represent 84% of the total carbon sequestration flows while covering only 56% of the total surface of the Mediterranean. However, heterogeneity and lack of information on carbon market prices as well as failure to recognize the co-benefits that these ecosystems generate have limited their integration into economic valuations. The absence of information on the real value of these ecosystems can lead to inefficient decision-making, often causing mismanagement (Canu et al., 2015). In addition, EEZs form the limit of the UNFCCC jurisdiction, leaving most of the open ocean and deep sea unconsidered with respect to climate mitigation and adaptation.

Moreover, the inclusion of social welfare variables in calculations of climate-change mitigating BCP impacts is now becoming key in policymaking related to open-ocean measures. Such regulations can also aim at minimizing the social impacts of carbon release. Hazards related to poor management of carbon sequestering ecosystems have both market and non-market consequences which need to be incorporated into cost-benefit weighing of Blue Carbon. The social cost of carbon (SCC) supports such cost-benefit evaluations of carbon emission mitigation policies. This process might be done using risk thresholds, independently from market-based economic impacts (Metcalf and Stock, 2017). Integrated Assessment models (IAMs), which so far only assess the risk of sea level rise, could present doubled levels of SCC with the inclusion of carbon-relevant ocean-related risks (Narita et al., 2020). SCC is included in the 5 "Shared Socioeconomic Pathways" (SSPs), scenarios weaved by outcomes of climate change, but needs to be adapted to SSPs which have a higher probability of occurrence (Yang et al., 2018).

Deep Sea Carbon

Only a handful of papers have considered the economic value of changes in ecosystem services in deep waters. The downward carbon flux at 1,000 m in the North Atlantic is projected to decrease by 27–43% under RCP 8.5 by 2,100; in the North Atlantic this is estimated to represent a loss of US $170–3,000 billion in abatement (mitigation) costs and US $23–401 billion in social costs (Barange et al., 2017). Others have highlighted the declining value of open ocean carbon sequestration in the eastern tropical Pacific (Martin et al., 2016) and, again, the Mediterranean (Canu et al., 2015). No economic estimates have been done in the direct context of ocean acidification in the deep sea. Surface waters are rapidly transported into the deep ocean and CO_2 is definitely rising, with carbonate saturation state declining in some deep waters such as in the Arctic and North Atlantic (Gehlen et al., 2014; Sweetman et al., 2017; Sulpis et al., 2018; Bindoff et al., 2019; FAO, 2019), but the consequences for ecosystems and their services are poorly known (Bindoff et al., 2019). Ocean deoxygenation will likely impair fisheries resources (Rose et al., 2019) with some of the greatest effects occurring at bathyal depths where oxygen minimum zones are expanding. Impacts of climate change are also expected to enhance carbon sequestration in key areas such as the Arctic and the Antarctic, where the decrease in ice will provide vast areas for carbon capture and longer blooms allow increasing sequestration in the Southern Ocean (Barnes and Tarling, 2017; Bax et al., 2020).

Shifting Our Markets

The Role of Natural Capital Valuation

Only an estimated 3% of current climate finance is allocated to nature-based solutions (NbS) (Raghav et al., 2020). Attracting the attention of the private sector provides opportunities to create sustainable business models that include social responsibility in order to invest in current carbon storage to comply with Paris agreements. Similarly, investment in natural climate solutions can support local communities and their ecosystems alike. Within carbon markets, buyers of Blue Carbon credits could choose to finance projects in areas that support their supply chain. For example, seafood companies might invest in the protection or restoration of Blue Carbon ecosystems for reasons beyond carbon offsets, such as co-benefits like nursery habitat protection of harvested species, protection from extreme events, mitigation of erosion and salinization, and improvement of human livelihoods (Vanderklift et al., 2019). In a situation of uncertainty, such approaches can be called low- or no-regret, as their cost is relatively low and they would provide benefits with or without expected impacts of climate change (Gattuso et al., 2018). The no-regret approach can be used at different social levels (households, communities, and local, national and international institutions) in order to increase resilience of social, economic and environmental policy benefits. However, these frameworks and their associated benefits are often poorly quantified, limiting investment. Providing information on the associated benefits of Blue Carbon can help increase financing for climate action while new resources flow in Siegel and Jorgensen (2011) and Ullman et al. (2013).

Protection and restoration of highly productive Blue Carbon coastal ecosystems may guarantee first, the integrity of carbon storage and second, the long-term removal of greenhouse gases from the atmosphere. It still remains challenging to trace the carbon sequestered back to its source and enhanced sequestration at the sink site needs to be assessed using management actions in source habitats (such as macroalgae). This cost-benefit approach common in financial investments can be used here to protect, invest in, and ultimately put these ecosystems on a sustainable path. But while the cost of conservation is well-understood and readily quantified, understanding the benefits of a vibrant natural world to our health and economic well-being depends on being able to show how natural resources, including species, habitats, and biodiversity, provide tangible value to humans (Chami et al., 2020a; Dasgupta, 2021). If we can reliably identify and measure the market-value of all the services provided by natural resources—such as carbon sequestration, flood control, fisheries support and more—we can then compare the present monetary value of these benefits with the cost of investing in them, just as we do for other financial assets. An example of such an approach is provided by Chami et al. (2019, 2020a) who apply financial valuation methods to value market services provided by cetaceans such as great whales in Chile and Brazil. Using existing markets and prices they derive an average lifetime value of $2 million per whale from carbon sequestration, whale tourism as well as impact on fisheries services. Their financial estimations use a growth model (logistic model in the case of cetaceans) and rely on the EU ETS to pin down the carbon price of $24.42 in 2019. By conserving these species, there is a potential to maintain an ongoing carbon sequestration pathway which is potentially vast compared to other designated ecosystems. Conservation efforts targeting these species can thus be fuelled by the carbon sequestration cycle in each individual along with their natural market services.

What Does the Valuation of Natural Capital Serve?

The resulting valuations can be quite effective at motivating environmental investment for several reasons. First, they show exactly what concrete services society currently receives from our stock of natural resources, which helps the public understand the relevance of these resources for its daily life. In addition, expressing the benefits of preserving natural resources in monetary terms allows for a dollar-to-dollar cost-benefit comparison, which is important as people are more comfortable making decisions when the stakes are expressed in financial terms. And finally, the value embodied in these natural assets can be very large—not only justifying the cost of preserving them, but also causing surprise and capturing the imagination of people who learn about the valuations. Behavioral economics research shows that people are more likely to purchase products or make investments that inspire these feelings (Chami et al., 2020b).

Valuing the benefits of ecosystem services highlights the cost of *doing nothing*, or *no-action*, related to degradation of ecosystems. In an estimation of global ES, Costanza et al. (2014) point out that the loss of ES between 1997 and 2011 due to land change lies between $4.3 and $20.2 trillion/year in 2007 $US. The core of restoration programs lies in our ability to quantify the value of Blue Carbon sequestration–also considering large uncertainty about methane emissions (Rosentreter and Williamson, 2020)–which would lead to the acquisition of carbon credits by countries investing in the calculated restoration of specific coastal ecosystems with high carbon potential (Kroeger et al., 2017; Tang et al., 2018).

Climate change, which jeopardizes the future and well-being of entire populations (Thiébault and Moatti, 2016), must be tackled through the combination of several approaches: (1) having a clear understanding of stakeholders' responsibilities for the management and governance of sensible ecosystems (either local community, industries, governmental institutions and international frameworks such as Regional Fisheries Management Organizations and the Convention on Biological Diversity), and (2) outlining a clear methodology to value carbon stocks, so as to identify key ecosystems which must be either restored or sustainably managed (Howard et al., 2017a,b). Scientific research can fuel those strategies as we enter the Decade for Ocean Science for Sustainable Development upon which future mitigation actions are likely to rely (IOC-R, 2021). Sustainable management of carbon sinks (and the local communities which can benefit from them) will also help nations meet their climate mitigation commitments. These include pledges taken under the umbrella of Nationally Appropriate Mitigations Actions (NAMAS), especially relevant in rapidly evolving coastal ecosystems that are vulnerable to abiotic variations such as temperature increase, extreme events (e.g., marine heatwaves), and sea-level rise (Laffoley and Grimsditch, 2009; Kirwan and Mudd, 2012; Gallo et al., 2017; Strydom et al., 2020). A stricter approach to climate change mitigation using the full potential of Blue Carbon across a range of ecosystems will rely on international cooperation, as suggested by the G20 Task Force 2 Policy Brief by Mansouri et al. (2020).

Governance

Financing

According to OECD (2016), a Blue economy would be able to create $3 trillion annually in gross value added by 2030. Blue growth is based on protection, conservation and investment in Blue natural capital, which, in turn, would lead to economic growth, but some aspects include highly destructive and climate-impairing practices like oil and gas extraction, or seabed mining. Cziesielski et al. (2021), challenged the traditional model of business-as-usual in which all sectors are interconnected but mainly related to the environment through exploitation. In their economic model, the environment (sustainably managed) was placed as the focal point of strategic development and other sectors (industry, communication, education, etc.) as secondary branching points. This method shows that the success and failure of each sector directly influences and impacts the environment, along with itself and others; in other words, the environmental sector is effectively "*the core of social and economic wealth.*" In the ocean's case, the High Level Panel for a Sustainable Ocean Economy (Ocean Panel) showed that 1 USD invested in actions linked with a Blue economy could generate 5 USD in global benefits (Konar and Ding, 2020). Then a key question must be formulated: How do we find this metaphorical "first" dollar, and thus initiate a productive and virtuous circle?

Direct and indirect public financing from international and regional organizations, states and local authorities can take the form of subsidies, grants, loans and transfers (international public aid) (Levallois, 2020). However, these modalities are mainly coming from general fiscal budgets, for which matters depend on public decision and are highly vulnerable to discretion and competing priorities of decision-makers. In this regard, the panorama of specific *affected* taxes going to ocean conservation still have extensive margins for improvement. Some examples exist but would need to be generalized to predict a scaled impact. For instance, in France, the owners of leisure boats are paying a yearly tax called "francization," related to the "Conservatoire du Littoral, 2020" for protecting coastlands (buying vulnerable lands, implementing ecologic restoration, projects against erosion, etc.). This tax is paid by more than 7-meter long boats, which are relevant targets related to coastal impacts (i.e., source of disturbance for *Posidonia* carbon sinks for instance), with low elasticity and a significant capacity to pay compared to the general population. This system gathers around 37.5 million euros/year, representing 72% of the institution's annual budget (51.7 million in 2020, according to the annual report of the "Conservatoire du Litoral").

It might be relevant to replicate these financial mechanisms for marine carbon conservation (both open ocean and deep sea) along with other targets. For instance, part of the product of existing port taxes could be dedicated to Blue economy investments. Even at low rates, the spectacular increase in sea traffic in the past decades and its associated impacts present increasingly important opportunities for financing (cruises, commercial, transport, tankers, etc.). In other terms, the reduction of the added value of highly pollutant activities allows for reinvestment in the Blue economy, "re"-creating sustainable added value. In any case, this framework would call for coordination between ports, cities and nations, through a highly competitive economy. A regional approach would be relevant to pilot activities before upscaling (for instance European ports in the Mediterranean Sea). Market-based responses and private financing (compensation of private companies, financial markets) can also be incentives to develop private "voluntary" participation in such investment efforts.

For example, carbon markets have arisen along with carbon pricing, a global trend triggered by increase in atmospheric CO_2 and in the destruction of carbon sinks. As Boyce (2018) explains, carbon pricing can be a challenging exercise: "carbon pricing initiatives around the world today cover approximately 8 gigatons of carbon dioxide emissions, equivalent to about 20% of global fossil energy fuel emissions and 15% of total CO_2 equivalent greenhouse gas emissions." Moreover, these markets are highly volatile, which poses a challenge for small-scale projects relying on coastal carbon returns. Hence, carbon pricing can either be through Emissions Trading Schemes (ETS), which facilitate the trade of permits for greenhouse gas emissions by capping the total level of emissions allowed; or carbon taxes, which set a price on carbon itself (The World Bank, 2019). Cabon pricing discourages emissions (Fankhauser and an Jotzo, 2018) and gives incentives to households, firms and governments to choose a more cost-effective way to reduce emissions (Boyce, 2018).

Carbon emissions from the degradation of coastal ecosystems such as mangroves, seagrasses, and saltmarshes, however, are rarely included in emissions accounting or carbon regular markets and protocols. In order to promote Blue Carbon projects in regulated carbon markets, reliable financial analyses must be taken into account to estimate Blue Carbon offsets, along with predictions of survival rates of new or restored vegetation in Blue Carbon ecosystems and measures of additional risks and benefits that could impede or enhance income flows (positive and negative externalities). In other words, a reliable scientific analysis of the permanence of these ecosystems must be carried out in order to guarantee that carbon is sequestered for long periods of time (from 25 to 100 years), without ecosystem degradation (Thamo and Pannell, 2016). These scientific efforts include research on natural sequestration and degradation of Blue Carbon ecosystems, impacts of human activity on the carbon cycle, exchanges of carbon between terrestrial and ocean ecosystems, the development and advocacy of sustainable policies, the creation of protection protocols as well as economic analyses of Blue Carbon impacts (IOC-R, 2021). Such an all-rounded approach can support existing and emerging carbon markets while calling for key ecosystem conservation and atmospheric carbon reduction. Limiting loss of coastal ecosystems may be more beneficial than implementing extensive restoration efforts in regions with lower carbon benefits (Pendleton et al., 2012; Vanderklift et al., 2019). This is particularly relevant as the carbon sequestration potential of coastal ecosystems such as mangroves is often underestimated since it is based on measuring methods used for terrestrial ecosystems which do not account for root and soil sequestration potential. A recent valuation project in Cispatá, Colombia, which covers 29,000 acres of mangroves aims to include the entirety of the carbon potential of this ecosystem as carbon credits which would differ from usual forestry credits both in their pricing potential and their management requirements for a reliable market (Klein, 2021).

Management

Strategies for sustainable management of Blue Carbon ecosystems can be based on their potential for adaptation and/or carbon sequestration. Adaptation plans are based on other services provided by Blue Carbon ecosystems such as protection from storm damage and flooding and the provision of resources, and the co-benefits that arise from them. Nationally Determined Contributions (NDCs) under the Paris Agreement provide a framework for declaring climate change mitigation intent which must be revised and enhanced every 5 years. Ocean adaptation is seen more clearly in the initial NDCs across the globe than carbon sequestration (Gallo et al., 2017): adaptation plans are found in 39% of parties with coastal wetland ecosystems whereas sequestration specific to coastal ecosystems is currently incorporated in the NDCs of only 19% of the relevant parties (Herr and Landis, 2016). Synergy between climate adaptation and mitigation strategies within NDCs provides an optimal approach for Blue Carbon ecosystem management. With the inclusion of carbon sequestration as an economic asset, well-rounded management planning for marine resources can benefit both a country's carbon emission mitigation strategy and its economic framework.

Recognition of Blue Carbon benefits are growing in the ocean policy community. Of the 47 submissions to the UNFCCC Ocean and Climate Change Dialogue held Dec. (2020) 85% of these referenced Blue Carbon, with an average of 7 mentions per submission that did, and greater intent among non-state submissions (89%) than state submissions (80%)[5]. The Blue Carbon initiative, coordinated by Conservation International, the Intergovernmental Oceanographic Commission and IUCN (and supported by the International Blue Carbon Scientific Working Group), strives to build Blue Carbon into NDCs, REDD+ (Reducing Emissions from Deforestation and forest Degradation), National Management plans and financing mechanisms such as the Green Climate Fund.

Blue Carbon mitigation efforts benefit from being incorporated into national strategies. In Madagascar, voluntary carbon markets, known to be better adapted for smaller-scale projects, are used across several project sites[6]. While legal regulations are being put into place to integrate coastal area management into national biodiversity plans (Decree N°2010-137) (Commission Nationale de Gestion intégrée des mangroves) (Herr et al., 2017). International policies such as REDD+, a program that compensates landowners for evident reductions in forest-based carbon emissions, incentivize the integration of coastal ecosystem restoration plans to a country's national climate change mitigation strategy. However, the incorporation of Blue Carbon plans into REDD+ activities will depend on a country's definition of forest ecosystems (Herr and Landis, 2016). Similarly, wetland inclusion is encouraged but not mandatory in the 2006 IPCC guidelines for National Greenhouse Gas Inventories despite the advantages of including coastal ecosystems in a country's greenhouse gases inventories[7]. Thus, heterogeneity in NDC commitments to Blue Carbon management has made worldwide policymaking challenging (Laurans et al., 2016). As explained by the Blue Carbon and NDC Guidelines developed by the Blue Carbon initiative[8], ecosystems can benefit NDCs by adding to the national green house gas emission mitigation strategies, providing ecosystem services to benefits local areas and communities, fuelling NDC achievements during the 5-year period, encouraging cross-sectorial efforts on the coasts to reach NDC goals (particularly in NDCs which incorporate Coastal Zone Management) and supporting Sustainable Blue Economies.

NDC inclusion of coastal ecosystems also encourages external financial support and climate finance for their sustainable management[2]. Coastal ecosystems need to be a part of a country's economic framework. A distinction among ecosystems during valuation calculation is important because certain habitats (e.g., seagrass meadows) can be expensive systems to restore (Bayraktarov et al., 2016). On the other hand, some vegetated coastal habitats have adaptive abilities and restoration capacities that could be invested in at minimal costs (Gedan et al., 2011; Duarte et al., 2013). This allows for strategies that relate to the maintenance of particular ecosystems such as the removal of anthropogenic nutrients in coastal habitats, the control of bioperturbator populations and the restoration of hydrology to increase carbon accumulation (Macreadie et al., 2017). These strategies include land-management regulations such as "other effective area-based conservation measures" (OECMs) which co-benefit biodiversity conservation under the umbrella of sustainable area management (Kalinina et al., 2021). Such restoration programs can also offset sequestered carbon losses due to damages, if properly managed, via processes of ocean zoning and marine spatial planning (Irving et al., 2011). Greiner et al. (2013) explain in their evaluation of the role of seagrass *Zostera marina* that restored meadows could accumulate quantities of carbon similar to a pristine one, given about a decade. Their 2011 social cost estimates are ∼7,000$ per year of carbon storage provided by restored meadows (Greiner et al., 2013). Hejnowicz et al. (2015) however arrived at higher estimates, using the cost of planting, monitoring, contracting and government oversight over the long term.

Despite recent developments in international policies regarding the protection of marine resources for climate change mitigation, results will only be achieved if there is ownership by

[5]Dobush, B.-J., Gallo, N. D., Guerra, M., Guilloux, B., Holland, E., Seabrook, S., et al. A new way forward for ocean climate policy as reflected in the unfccc ocean and climate change dialogue submissions. climate policy. (in revision).
[6]see https://blueventures.org/conservation/blue-forests/.

[7]https://www.ipcc.ch/report/2006-ipcc-guidelines-for-national-greenhouse-gas-inventories/
[8]See https://www.thebluecarboninitiative.org.

local actors. Legal and territorial planning frameworks, including social responsibility, must be defined on coastal compensation sites (Thamo and Pannell, 2016). This territorial perspective on governance might integrate potentially contradictory points of views but must define joint common responses. Examples include the 2020 Earth Security report on Mangrove financing, presenting a case for investment in mangrove restoration using a network of 40 key cities which can focus on mangrove protection while carbon markets evolve to reach global importance. This requires integration of information and data across institutions (on national levels as well as local representation of ministries), along with the coordination of local authorities and participation of communities and economic stakeholders in public decision. Adequate realistic planning involving local partners will facilitate measuring benefits and define priorities for conservation and/or restoration, and implement concrete activities to achieve results… *if sustainable financing is availab*le. On the one hand, financial incentives for carbon compensation are able to attract short-term funds but are not necessarily adapted to local expectations and remain lacking in local knowledge and input. Territorial planning with a community perspective can be adapted to local needs, becoming an efficient source of long-term development, but lacking in economic power. These often-opposing approaches would warrant the role of public institutions to foster dialogue and guarantee sustainability, in order to develop complementary practices. Involving local populations in governance of ecosystem management plans allows for Blue Carbon to support a steady resource flow into local communities beyond current fragmented financing plans.

In open-ocean ecosystems, management of carbon stocks is also challenged by the movement of carbon across the water column and national and jurisdictional borders (Luisetti et al., 2020). Though occurring naturally, these movements can also be triggered by anthropogenic activities indiscriminately impacting sediment delivery in the water (Crooks et al., 2018). Little is known about valuation of carbon as a transboundary resource and the uncertainty regarding the origin of the carbon makes its valuation challenging. Carbon frameworks in international waters (covering about 60% of the oceans) are just beginning to take form, mostly relying on voluntary commitments for management. The United Nations Convention on the Law of the Sea (UNCLOS) of 10th December 1982 mentions that “No State may validly purport to subject any part of the high seas to its sovereignty” (article n°89). Defined by the parts of the ocean not included in EEZ, territorial, internal or archipelagic water, this legal criterion of high seas represents about 64% of the surface of the ocean. This is also the case for seabed (“area”) and its resources, which are considered as “common heritage of mankind” (article 136) [United Nations Convention on the Law of the Sea (UNCLOS), 1982]. Activities in such areas are under the control of the International Seabed Authority and shall take into account the principle of protection and conservation of natural resources. This question calls for international innovative coordination and joint strategies, in order to avoid habitat degradation that releases carbon and ensure the integrity of carbon cycling and sequestration. The potential international support granted to Blue Carbon rich countries is one of many incentives arising with cross-sectorial carbon management (Chan, 2021). Other initiatives include the UNEP Regional Seas Program which proposes a shared areas approach based on tackling “ocean grabbing,” an issue of powerful stakeholders attempting to secure other resources from the ocean along with carbon rights (Barbesgaard, 2018). Climate change considerations and carbon conservation could be addressed by elements of the ongoing Marine Biodiversity of Areas Beyond National Jurisdiction (BBNJ) treaty negotiations such as area-based management tools (including MPAs), and environmental impact assessment, but climate issues have not risen to a high priority in this historic negotiation (Tessnow-von Wysocki and Vadrot, 2020).

The modernization of joint coordination strategies among different states, and for international waters, among different UN agencies, is necessary in order to define achievable, realistic, and progressive protocols agreements. These would not be sectoral but comprehensive, with an integrated management perspective (involving coastal ecosystems, open water and deep-sea ecosystems with environmental, social and economic questions). Even the UN Conference of Parties (COP) has been shown to benefit from restructuring approaches in order to adapt to the evolving climate crisis and the stakeholders impacted by and involved with its outcomes (Ferrer et al., 2021). Coherency is mandatory to achieve long-term results across domains of fishery management, biodiversity conservation, transport and tourism policing. It is indispensable to foster real management plans from a local level (including methods of ecosystem engineering, the ecological enhancement of marine Blue infrastructure) to an international level (development of MPAs, marine spatial planning (MSP), transnational protected areas and creation of cross-border online platforms for carbon and biodiversity offsets). Platform such as the Ocean and Climate Change dialogue support discussions which can strengthen a common understanding of the gravity of the situation across stakeholders and scale of impact (UNFCCC, 2021). Sustainable management of transboundary marine resources through integrated approaches presents a unique opportunity to avoid conflicts and develop cooperation for shared benefits. Transboundary water pollution and climate change are key areas for improvement (Giupponi and Gain, 2017). Marine ecosystems have no borders and acknowledging connectivity between ecosystems is essential to sustainably manage and develop marine resources to their maximum potential. Thus, marine management can be an opportunity to develop cooperation. Potential approaches include leaders’ training programs with the goal to increase awareness with clear and precise communication about the value of Blue Natural Capital, along with clear transboundary management strategies for marine resources with a transboundary diagnostic analysis, and pertinent and achievable strategic action programs (Cziesielski et al., 2021).

Environmental measures should tackle, both, terrestrial and marine ecosystems, with one as a continuum of the other. Coral reef restoration can increase coastal resilience to sea level rise and flooding and provide valuable environmental services for local populations (Hamerkop, 2021). From the opposite direction, water pollution in rivers contributes to ocean ecosystem degradation, via eutrophication and the formation of dead zones. River basin management planning (and associated financial mechanisms) must integrate the relationship between freshwater resources and marine ecosystems.

Holistic and integrative concepts are key to account for differences in representation. Implementing natural based solutions (NbS), for which Blue Carbon is one type, could support the integration of this continuum between mitigation/adaptation and terrestrial/marine solutions. NbS can make impactful changes if built for the long-term and continuously measured with the right metrics to support a range of ecosystems and their local communities' rights and needs (Girardin et al., 2021). The success of NbS implementation will depend on close collaboration between a wide range of stakeholders and polycentric governance structure as well as on the clarification of values and interests they have in common (Martin et al., 2021). Outlining specific policy targets for NbS throughout the project duration can strengthen the effectiveness of these strategies (OECD, 2021). This collaborative approach would include citizens, partnerships with environmental organizations and universities, private and public sector and community action and engagement (Cohen-Shacham et al., 2016).

Local Populations

As mentioned above, local actors are key to the long-term success and development of Blue Carbon ecosystem management schemes, particularly as 120 million people worldwide now live near mangroves (UNEP, 2014). According to the International Federation of the Red Cross and Red Crescent the replanting of mangroves in Vietnam between 1998 and 2002 has reduced the cost of dyke maintenance by $7.3 million a year for an investment of $1.1 million (International Federation of Red Cross Red Crescent Societies, 2002). Moreover, the Markets and Mangrove project directly links income for the community and mangrove ES, as shrimp farmers now have financial incentives to protect mangroves, with the assurance of higher revenues from their newly obtained organic certification (McEwin and McNally, 2014; Wylie et al., 2016). Similarly, the mangrove restoration program Mikoko Pamoja[9] in Kenya uses sales from carbon credits to support schooling and the provision of piped water in the community (Wylie et al., 2016). In Indonesia, 13 million metric tons of Blue Carbon have been stopped from release into the atmosphere between 2000 and 2010 due to protected coastal areas, amounting to $540 million in social welfare benefits (as calculated by Pendleton et al., 2012; Miteva et al., 2015).

Socio Manglar (Ecuador), a program in which communities receive cash directly for their sustainable management of mangrove forests (Herr et al., 2017) demonstrates that mitigation plans ought to aim for local involvement and governance as an ultimate objective. As previously stated, ES, including carbon sequestration, have a number of costs and benefits which are likely better defined by local populations (Bennett, 2014). However, they have been vastly excluded from decision-making processes so far (Hejnowicz et al., 2015). Ownership and accountability are often an obstacle to community-led governance but can be mitigated by introducing external parties such as research institutes to oversee operations (Vanderklift et al., 2019). Wider scale policies, however, risk the exclusion of local communities from decision-making and governance, a better understanding of the social impact of Blue Carbon has proven widely successful in specific localizations so far.

Time presents a challenge for these integrative processes: ecosystem service assessments must be adapted to the timeline of policy decision-making (Ruckelshaus et al., 2015). Overall, successful mitigation plans rely on science to tackle knowledge gaps with branching approaches such as the Integrated Ocean Carbon research (IOC-R, 2021), funding from investors whose interests are well-understood (Vanderklift et al., 2019), financial processes that support local communities, transparency in decision-making, cross-sectorial support and their incorporation in national efforts for climate change mitigation. The "Livelihoods Carbon Fund" in Senegal presents a strong example of such a cyclic plan: using funding from private companies to offset their carbon impact, this initiative started in 2009 supported the planting of 80 million mangroves, which are impacting not only the carbon sequestration potential of the area but also biodiversity (fish, shrimp and oysters mainly) and farming activities, by counteracting salinization of rice fields[10]. These sustainable long-term impacts support poverty alleviation while protecting key ecosystems and mitigating climate change. Thus, sustainable biodiversity and ecosystem management can provide a foundation upon which to build strategies for poverty alleviation and sustainable community maintenance and growth (Bawa et al., 2020).

In order to meet the Paris Agreement and Sustainable Development Goals 14 (conserve and sustainably use the oceans, seas and marine resources) and 13 (take urgent measures to combat climate change and its impacts), Marine Spatial planning (MSP) not only focuses on reducing carbon emissions but must also contribute to net zero commitments of a country. MSP also generates collateral benefits such as the promotion of gender equality, solid and more sustainable rural livelihoods and production of new jobs (among others). These systemic co-benefits promote public participation, information sharing and dissemination in order to raise awareness of climate justice. Furthermore, MSP is a platform upon which local populations can develop a direct line of contact with government institutions, local authorities and the private sector with the aim to preserve marine ecosystems and reap economic benefits. This opportunity empowers communities, which can build their capacity to shift natural sourcing practices toward more sustainable paradigms that could be achieved due to increased awareness and education programs (Cantral et al., 2012; Suzanne et al., 2019).

[9]program Mikoko Pamoja See https://www.mangrovealliance.org/mikoko-pamoja/

[10]The "Livelihoods Carbon Fund" in Senegal See https://livelihoods.eu/portfolio/oceanium-senegal/

Marine Protected Areas

The implementation of Marine Protected Areas (MPAs) has risen over the last decades as a promising option to mitigate climate change impacts on carbon removal processes, as long as regulation strategies guaranty the integrity of natural carbon stores (Jones et al., 2018; Bindoff et al., 2019). MPAs have emerged as important governance responses to coordinate ecosystem management, resource utilization and biodiversity conservation, and they currently represent 5.3 % of the global ocean and 1.2% of high seas[11]. MPAs offer more financial stability than carbon markets by securing resource supply and stable regulations (Thomas et al., 2010). These are achieved through cross-sectorial efforts and agreements on jurisdiction and accountability (Howard et al., 2017a). This reliability makes MPAs sustainably beneficial to a country's overall GHG emissions accounting, providing further incentives for their conservation. Protection of the carbon services provided by the coastal ecosystems remains challenged by governance boundaries (such as the UNFCCC) and competing societal needs. MPA design rarely incorporates carbon services, and marine MPAs currently cover and fully protect <3% of the oceans[12]. However, 50 World Heritage Sites currently cover 21% of the global area of documented Blue Carbon ecosystems (29% of seagrass, 7.2% of tidal marsh and 8–9% of mangrove forest) (UNESCO, 2020). In addition, some ecosystem services provided by open-ocean ecosystems are not yet replaceable by human industries, highlighting the importance of protection. Policymaking has focused on coastal ecosystems such as mangrove forests, salt marshes and seagrass meadows, disregarding the potential of open-ocean and deep-sea ecosystems to provide support to mitigation efforts. Scotland has begun to view the potential of Blue Carbon as an incentive in its own right for the implementation of MPAs so it can be directly considered by marine management, both on regional and national scales (Laffoley, 2020).

Incentives for the protection of key areas include other ecosystem services provided as well as social benefits derived from ocean protection inclusion in national policies. MPAs continue to present potential for wider protection of key areas, facilitating governance issues and financing opportunities; but there is a need for international frameworks to step in in order to account for the movement of carbon through national borders and to facilitate cost-effectiveness and economic accountability of ocean-based measures. "Other effective area-based conservation measures" (OECMs) can also serve conservation purposes if properly managed while their main aim focuses on sustainable land management (Kalinina et al., 2021). In both cases, Blue Carbon needs to be an incentive for ecosystem protection in its own right, both recognized and sustainably considered by marine management regulations.

Deep-sea ecosystems offer potential for long-term carbon storage in stable conditions, offering yet another path toward climate change mitigation in open waters, along with an array of other ecosystem services. However, the complexity of deep-sea carbon storage presents two main challenges: (1) data regarding the level of risk which can be sustained by these ecosystems and the practical economic valuation of their carbon services as opposed to emission risks is lacking and (2) these ecosystems are currently vulnerable to anthropogenic disturbance but MPAs can alleviate this risk and contribute to sustainable management practices. There is in any case a need for innovative legal and governance frameworks in order to respond to the challenges of Blue Carbon in deep-sea ecosystems, according to the legal statute of "high sea" and "area" (seabed), beyond national jurisdictions.

[11] See http://www.mpatlas.org.
[12] https://mpatlas.org

CONCLUSION

The Paris Agreement requires serious commitment at a country and industry level to achieve carbon neutrality by 2050 if the world is to avoid breaching the 1.5–2.0 degree ceiling. This looming deadline places demands onto all stakeholders to neutralize and offset carbon emissions. Among potential answers are nature-based solutions, which play a key role in maintaining active carbon sequestration processes and preventing human assisted-nature-based emissions (e.g., from habitat loss and degradation). Because they are both accessible and co-beneficial for local communities, blue nature-based solutions should be paired with dramatically increased efforts to reduce GHG emissions. Similar to terrestrial carbon in forests, the ocean captures carbon in a range of ecosystems (coastal, deep sea and open ocean) which often offer other services with shared benefits across the society.

This paper suggests how these various ocean ecosystems could support mitigation strategies and carbon stock conservation when sustainably managed across sectors. Financing from stakeholders, who would benefit from ecosystem services as well as from carbon credits, requires a transparent and credible system for managing such a market. Although "climate bandwagoning" (Chan, 2021) commonly justifies political and economic action under a cover of climate mitigation, it now needs to translate into applicable and fast-paced governance practices and policies, starting with partnerships with the private sector as well as the expansion of the tax base. These conservation efforts can only succeed if local communities are part of the decision-making process, where they stand to directly benefit from the meaningful employment and steady income that would help ensure ownership of these efforts.

The COVID-19 crisis has clearly demonstrated the consequences of poor management of the natural world. Many believe that COP26 in November 2021 is the best last chance to get the climate change risk under control. We argue that ocean solutions are a key part of the mix and hope that the protection and restoration of marine carbon stocks and sequestration processes will be part of the COP26 discussions since this will also help address the marine biodiversity crisis and reduce risks of impacts to critical ocean system functions. The post-pandemic period presents an opportunity to reboot our paths to economic development by taking into account the potential and the value of ocean system services, starting with integrated policies tied to economic, social and environmental recovery strategies. Global partnerships leading to immediate actions are needed to pair social protection with climate action and economic recovery, in order to rebuild and transform economies from an ecological standpoint.

AUTHOR CONTRIBUTIONS

NH and LL led the work. LL, MS, and JH-S contributed mainly in the ecological sections. NH, RC, LL, and MB contributed to the economic and policy sections. All the authors contributed to the manuscript and revised the final manuscript.

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Ocean Acidification in the Arctic in a Multi-Regulatory, Climate Justice Perspective

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The latest IPCC report on Ocean and Cryosphere in a Changing Climate, which builds upon previous IPCC's reports, established a causal link between anthropogenic impacts and ocean acidification, by noting a significant decrease in the Ocean's uptake of CO^2, with consequent damage to Earth's ecosystems, which in turn has traceable repercussions on the Arctic Ocean and then from the Arctic to the Planet Earth. The impact of ocean acidification is not only in the biological ecosystem but also on human activities, such as livelihood, food security, socio-economic security and developing communities. However, who can possibly be held ethically/legally responsible for ocean acidification from a climate justice perspective? Since what happens in the Arctic does not stay there, a more systematic law and policy approach to study options and responses in a multi-level, climate- ethical, global perceptive is needed. This paper sheds light on the legal responses available at global, regional and national levels to ocean acidification in a law of the sea and ocean context, both in the Arctic and from the Arctic. The gaps in legal and policy responses in connection to the ethical climate component will be identified. It will shed light on the planetary limits that humanity needs to stay within in order to maintain the future of the Earth. Since it touches upon questions of legal responsibility, on who is responsible for ocean acidification, it will connect to the "supply side" of fossil fuels production and global extraction projects causing anthropogenic CO^2 emissions, one of the major causes of ocean acidification. It will also identify which actors, be they "officials" or "non-officials" (such as international organizations, states, regional institutes, Arctic citizens or even *forums*) should be held ethically responsible, and who should take action.

Keywords: ocean acidification, climate change and ocean governance, arctic climate justice and ethics, legal responses to ocean acidification, Arctic Ocean acidification, environmental justice

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INTRODUCTION AND HYPOTHESES

Climate change is determining dramatic changes to ocean ecosystems. It poses threats to marine biodiversity and in turn to the entire human dimension associated with it, such as goods, services, livelihoods that the ocean provides. One of these threats is ocean acidification. The latest IPCC Report on Ocean and Cryosphere in a Changing Climate of 2019 established a link between anthropogenic impacts and ocean acidification noting a significant decrease in the Ocean's uptake of CO_2 with consequent damage to Earth's ecosystems, which in turn has traceable repercussions on the Arctic Ocean and from the Arctic to the Planet Earth. Ocean acidification is crucial because can be conceived as an important indicator of the nexus between climate change and oceans ecosystems under the threat of the current Anthropocene epoch we are living in. However, there is much uncertainty and little knowledge about what the responses should be from a governance perspective, including the role of international law in addressing ocean acidification as an *equality and justice* problem inherent to climate change rather than just a consequence thereof. This lack of knowledge could have repercussions on mitigation, adaptation and on determining who should be held responsible for ocean acidification.

There are major knowledge gaps in the current literature and weak insight and solutions from a social science perspective. This article intends to contribute to filling this gap by analyzing responses to ocean acidification in international law with special attention to Arctic Ocean acidification from a new angle combining governance and climate justice approaches. The hope is that by discussing whether ocean acidification should be treated as a threat of climate change rather than a concurrent problem will lend greater clarity to the issues in question. Other equally relevant hypotheses are, whether the current instruments of international law are fit to address this issue and if the connection between responses in international law to climate justice arguments could provide new avenues for increased responses such as for example, if there should be a *forum* in the Arctic Ocean to tackle ocean acidification through the combination of existing agreements and institutions.

The rifts in legal and policy responses in connection to ethical, climate justice components will be identified also by highlighting the planetary limits that humanity needs to stay within in order to maintain the future of the Earth. Since this perspective touches on questions of legal responsibility, liability and on who is actually responsible for ocean acidification it will connect to the supply side of fossil fuels production and global extraction projects causing CO_2 emissions as one of the main causes of ocean acidification. It will also individualize which actors should be ethically responsible and who should take action.

In the case of ocean acidification either at general level or more specifically in the Arctic Ocean Acidification context, it is unlikely that a single institution or level of governance or any single set of policies across institutions will be able to tackle the problem of ocean acidification since ocean acidification is a collective action problem occurring in a shared space of global commons. From this angle it is no longer *equitable* and *just* that governments alone should be responsible or finance ocean acidification, a problem the latter not involving only the public sectors but also the private sectors and not only official actors but non-official actors, all of which contribute to the large pollution problem. There are no incentives to change "business as usual" and there are several weaknesses in relying upon social institutions of governments, or insurance of civil liability for managing and transferring risks of ocean acidification.

However, theories of climate justice connected with principles of environmental law could suggest approaching the problem form a different angle. From a climate justice perspective damages caused by ocean acidification could be prevented by creating a *forum* at a regional level, which in the case of Arctic Ocean Acidification could be established at the Arctic Council (AC) level. A forum of this kind would be able to push for a behavioral change by instilling a new idea of "climate ethical ocean justice" vital to protecting the space of one of the most relevant planetary boundaries of the Earth system.

In order to describe this new angle and to unfold the problem of ocean acidification by operationalizing the hypotheses, in a multi-level holistic vision, this article is structured as follows: the next section sets the field of research by establishing a connection between ocean acidification in the Arctic and from the Arctic both as a threat and consequence of climate change (rather than a concurrent problem) and by explaining how the conceptualization of ocean acidification can be perceived as a planetary boundary within a space that preserves and guarantees an *equitable* and *just* future. The subsequent section depicts the multi-regulatory landscape that can be applied with special emphasis on Arctic Ocean Acidification, identifying the gaps in terms of responses with emphasis on the absence of a threshold or boundary line for Ocean Acidification that should not be surpassed in order to bring the level of Ocean Acidification to an acceptable level. In fact, an acidity threshold would be relevant in environmental treaty law in order to establish new standards and thresholds to be incorporated in future law and decision-making policy and legal instruments. The penultimate section connects ethical and climate justice arguments at the core of ocean acidification by establishing directs links between critical environmental theories and environmental law responses that could serve as a foundation for a new holistic governance approach. The concluding section summarizes the implication of the approach for future recommendation and policy perspectives.

OCEAN ACIDIFICATION IN THE ARCTIC AND FROM THE ARCTIC AS A CONSEQUENCE OF CLIMATE CHANGE

Oceans are the backbone of our planet and play a crucial role in regulating the impact of climate change and partly by absorbing excess of heat, and partly by acting as enormous sinks for carbon emissions. The Intergovernmental Panel on Climate Change (IPCC) assessed that 90 % of the combined heat stored in the climate systems has been absorbed by oceans between 1971 and 2010 [Intergovernmental Panel on Climate Change (IPCC), 2014]. The last IPCC Special Report on the Ocean and Cryosphere in a Changing Climate of 2019 established that oceans have absorbed approximately 30 % of emitted anthropogenic carbon dioxide [Intergovernmental Panel on Climate Change (IPCC), 2019] thus changing the ocean's chemistry and leading to ocean acidification, a process likely to have wide-ranging ramifications for marine biodiversity, biogeochemical processes, the goods and services derived from the oceans and the billions of people depending on it (Harrould-Kolieb, 2020).

The impact of ocean acidification is expected to include, but is not limited to, economic losses from a decline in fisheries and tourism, impacts on human health and decreased coastal protection [Arctic Monitoring and Assessment Programme (AMAP), 2018]. Ocean acidification is likely to cause major shifts in marine ecosystems, including the loss of most coral reefs globally and a decline of species globally (Eyre et al., 2018).

Geochemical information informs about potential risks to the Arctic and from the Arctic to the rest of the planet. In the Arctic Ocean, the cold surfaces waters absorb CO^2 more rapidly than warmer waters, leading to a disproportionately higher fraction of the global net CO^2 uptake and climate changes have intensified this susceptibility to ocean acidification [US Geological Survey (USGS), 2012]. Ocean acidification in the Arctic Ocean reduces shell formation, determines habitat loss and less food for predators, thus damaging ecosystems and ecosystem services (Scott et al., 2020). Polar waters will be the first to see the lowering of carbonate ion concentrations to such an extent that shell-forming organisms will not be able to calcify[1]. According to the IPCC Fifth Assessment Report, marine organisms are at risk from progressively lower oxygen levels and higher rates of ocean acidification. It underlines that coral reefs and Polar ecosystems are highly vulnerable[2].

Because of global warming, science proved the loss of Polar ice and over the last two decades, the Greenland and Antarctic ice sheets have been losing mass, glaciers have continued to shrink almost worldwide, and shrinking Arctic sea ice and Northern Hemisphere spring snow cover have contributed to a decrease in extent[3].

The Arctic Monitoring Assessment Programme (AMAP) found that ocean acidification, particularly coupled with ocean warming and deoxygenation, will drive changes in the marine ecosystems and impacts Arctic biota and it is likely that ocean acidification will drive changes at a magnitude that will affect the living resources in the Arctic and surrounding regions [Arctic Monitoring and Assessment Programme (AMAP), 2018].

The Earth system is undergoing human-induced changes in a dramatic scale and human interference with natural systems has caused a high level of *uncertainty* about what the planet will look like in the future (Lim, 2019). The concept of "planetary boundaries" describes the important interdependence of the major environmental challenges faced by the Earth System. Ocean acidification is one of the "planetary boundaries." Planetary boundaries have been defined as a series of biophysical boundaries at the planetary level (Rockström et al., 2009). They include issues such as climate change, biodiversity, freshwater use and ocean acidification. They function as a "safe operating space for humanity," i.e., denoting the planetary-scale limits that human activity needs to stay within in order to maintain the functioning of the Earth's systems in a manner which will allow continued human development (Rockström et al., 2009). Climate change is central to this anthropogenic disruption interlinked with each of the planetary boundaries, including ocean acidification (Minas, 2019). This explains why it is so crucial that law and policy treats the two in synergy: climate change and marine environmental protection together and not in disjunction.

The feedback effects of ocean warming and ocean acidification from the Arctic to the rest of the planet may itself aggravate climate change, and the melting of ice (Reid et al., 2009). Ice melting includes the melting of permafrost in the Arctic, resulting in the release of methane which is a green gas some 34 times more powerful than CO.[2] Occurring over a 100 years period, methane has been released from the seafloor, in turn exacerbating ocean acidification and ultimately entering the atmosphere (Brown et al., 2016). How can the law facilitate the preservation of the planetary boundaries including ocean acidification to facilitate an *equitable* and *just* future and preserve the planet in a safe space for our future generations?

The conceptualization of ocean acidification as a planetary boundary within a margin to guarantee to guarantee an *equitable* and *just* future for humans and our planet is still not fully understood. There is a chasm of ignorance and lack of regulation from a legal and ethical perspective on how ocean acidification should be conceived and factored in law, decision making associated with marine planning, fisheries management, and area-based protection under the law of the sea (Scott, 2020). Ocean acidification is barely present in environmental justice literature as well.

In the latest IPCC Special Report on Ocean & Cryosphere in a Changing Climate of 2019, the scientific assessment on ocean acidification notes that the effects of ocean acidification are geographically highly heterogeneous and uncertain but there is an improvement of the understanding of the natural science processes underpinning ocean acidification [Intergovernmental Panel on Climate Change (IPCC), 2019]. From an international law, governance and ethical perspective, ocean acidification has not been included in the Polar Chapter (Chapter 3) of the IPCC Special Report on Ocean and Cryosphere in a Changing Climate of 2019 due to the insufficient amount of literature (at the time of the writing the Report) necessary to elaborate a qualitative and quantitative scientific assessment [Intergovernmental Panel on Climate Change (IPCC), 2019]. Nor has ocean acidification been treated in the theories of environmental justice in connection to either general environmental justice literature, or to specific Arctic Environmental justice literature.

In fact, while the root cause of ocean acidification lies in human policies and behaviors driving society's dependence on fossil fuel, resulting in elevated CO_2, there is still a hole/reluctance in the social science literature, specifically in the law and policy area, to engage in ocean acidification (Jagers et al., 2019). Actually, there are still holes in our knowledge about which kind of regimes, policies, legal provisions, and mechanisms can address ocean acidification both in the Polar Regions and the rest of the planet. Provisions of treaty law and regional agreements are applicable but do not address directly the problems of ocean acidification (Oral, 2018). However, it is possible to map the legal regimes applicable to influence mitigation, adaptation and resilience of ocean acidification.

Yet, little is known on how about how society can respond to ocean acidification (Jagers et al., 2019) which is the main reason why possible human responses to ocean acidification can actually

[1]Antarctic and Climate Ecosystems Cooperative Research Centre, Position Analysis: CO_2 Emissions and Climate Change: Ocean impacts and Adaptation issues. (2008). 5, in Baird et al. (2009).

[2]IPCC Fifth Assessment Report (supra 1).

[3]IPCC Fifth Assessment Report (supra 1 and 5).

be considered as a gap of knowledge in term of human responses not only in Chapter 3 on Polar Issues but also in the entire Special Report on Ocean & Cryosphere in a Changing Climate of 2019 in connection to the other Chapters dealing with regions other than the Polar ones.

Ocean acidification is linked to *climate justice* and the *unequal distribution* of global pollution. Ocean acidification is a "collective action problem" of global pollution as there are reasons to believe that the practices from which an unequal distribution of global pollution effects emerge do not benefit every party and inequalities in the distribution of effects and such practices should not be permitted (Skillington, 2017a). Current practices of resource exploitation, especially in the Arctic, do not benefit all peoples but rather undermine the quality of life of a global majority.

LEGAL RESPONSES AT GLOBAL AND REGIONAL LEVEL

The legal landscape responding to ocean acidification is composed of a regime complex of multi-regulatory systems of sources of law and policy at global, regional and national level characterized by hard law, soft law, standards, and decision-making tools. The multi-layered and deformalized structure where different actors (both officials and non-officials) operates highlight the possible synergistic and interactive application of the different sources of law and policy to ocean acidification.

The three main regimes that hold prominence in ocean acidification classified in "global regimes and regional regimes" are from: (1) the climate change regimes, (2) the marine pollutions regimes, and (3) the biodiversity regimes. Ocean acidification is thus not regulated by one single regime but governed by a "regime complex"[4] where different sources of law and policy interact and overlap without coordination even though several mitigation strategies on ocean acidification have been initiated through multilateral cooperation (Jagers et al., 2019).

At the global level, despite the increasing knowledge about ocean acidification, there are no provisions explicitly aimed at regulating ocean acidification and no treaties combatting ocean acidification.

The most relevant frameworks existing at global level that are also applicable to the Arctic Ocean are the United Nations Convention on the Law of the Sea (UNCLOS)[5], the United Nations Convention on Climate Change (UNFCCC)[6], the Paris Agreement[7], the Biodiversity Convention (CBD) [8]although there are other relevant developments that can be considered such as for example within the typology of the atmospheric pollution global regimes.

Global Regimes

Amongst the main global pollution regimes, the 1982 United Nations Convention on the Law of the Sea (UNCLOS) is a legal framework within which all activities in the oceans and seas must be carried out. UNCLOS is of strategic importance for global, regional and national level action and cooperation in the marine sector. Within this framework, the Arctic Ocean can be seen as an ecosystem to be protected. Some environmental provisions of UNCLOS are relevant for ocean de-acidification but also with the view of applicability to the Arctic Ocean.

UNCLOS's part dedicated to environmental protection in general is Part XII where pollution of the marine environment has been defined in general terms in Article 1 (1) 4 which includes ocean acidification effects[9]. Art. 194 of UNCLOS provide that States shall take all measure necessary to prevent, reduce and control pollution of the marine environment from any source[10]. Paragraph 5 of Art. 194 is of particular relevance regarding ocean acidification in the Arctic as it deals with vulnerable areas[11].

UNCLOS requires States to take those measures necessary to prevent, reduce and control pollution of the marine environment and to adopt national laws and regulation to prevent and reduce pollution of the marine environment through the atmosphere. This includes the introduction by man of energy into the marine environment, which drives the increase in energy stored in the oceans and its associated impacts, including oceans warming, sea level rise, marine species redistribution, impacts on ecosystems. The notion of "pollution of the marine environment" therefore includes the direct introduction of anthropogenic carbon dioxide into the marine environment, a cause of ocean acidification.

Since there is no coordination and linkage between conservation and management measures and the impact of climate change as present, the UNCLOS convention is

[4]A "regime complex" is a collection of governance arrangements that are linked together in the sense that they address matters related to a common issue area or spatially defined region but that are not hierarchically related in the sense that they all fit within some well-defined institutional architecture. The theorists of this way of thinking about governance have focused on cases like the regime complex for plant genetic resources and the regime of climate change. For "complex regimes," see Oran (2012).

[5]United Nations Convention on the Law of the Sea (UNCLOS), 10 December 1982.

[6]United Nations Framework Convention on Climate Change, (UNFCCC), 9 May 1992.

[7]Paris Agreement on Climate Change, UN Doc. FCCC/CP2015/L.9/Rev.1, 12 December 2015.

[8]Convention on Biological Diversity (CBD), 5 June 1998.

[9]Art. 1 (1) 4 of UNCLOS states "Pollution of the marine environment means the introduction by man, directly or indirectly, of substances or energy into the marine environment, including estuaries, which results or is likely to result in such deleterious effects as harm to living resources and marine life, hazards to human health, hindrance to marine activities including fishing and other legitimate use of the sea, impairment of quality for use of sea water and reduction amenities".

[10]Art. 194 of UNCLOS specify that states shall take all measures necessary to prevent, reduce and control pollution of the marine environment from any source. Sources of pollution are for example: pollution from land-based sources, pollution from or through the atmosphere, pollution by dumping, pollution from vessels, pollution from seabed activities, pollution from other installations and services operating in the marine environment and pollution from activities in the Area. With the term "measures" contained in this article, it is understood that it is inclusive of those necessary to protect and preserve rare or fragile ecosystems as well as the habitat of depleted, threatened or endangered species and other forms of marine life.

[11]Article 194 (5) states "The measures taken in accordance with this Part shall include those necessary to protect and preserve rare or fragile ecosystems as well as the habitat of depleted, threatened or endangered species and other forms of marine life".

supplemented by more detailed regimes, including those regulating dumping at sea and land and atmospheric source marine pollution both of which are applicable to ocean acidification and to some degree to the Arctic Ocean. An example is the 1972 London Convention[12] and its 1996 London Protocol[13], negotiated to replace the 1972 London Convention both aiming at preventing the pollution of the sea by the dumping of waste or other matters liable to create hazards to human health, and harm living resources, including marine life. The convention also applies to the dumping of active waste in parts of the Arctic Ocean.

With regards to the *global climate regimes*, some of the most relevant regimes dealing with ocean acidification are the United Nations Framework Convention on Climate Change (UNFCCC), and the Kyoto Protocol (Kyoto Protocol), recently replaced in 2020 by the Paris Agreement of 2015. The UNFCCC convention has clear implications for the Arctic Ocean as the effects of global warming could have a devastating impact upon various types of ice. Not only would the release of fresh water from the ice cap increase a rise in the sea level but it would also have an impact on the marine ecosystem. The Arctic States[14] are large, industrial states, which have significant temperate lands in addition to their Polar claims and interests. The problem of climate change is truly global, one of which states need to cooperate collectively. All these climate change regimes focus on reducing the Greenhouse Gas Emissions (GHGs) that cause ocean acidification. However, ocean acidification had not been examined scientifically in detail at the time of negotiations of these two treaties—the UNFCCC and the Kyoto Protocol. There is no mention of ocean acidification in either of them. Nor is any mention of the problem of ocean acidification in the recent Paris Agreement.

Article 2 of the UNFCCC is relevant for ocean acidification stating that the object is the "*stabilization of greenhouse gas emissions in the atmosphere at a level that would prevent dangerous anthropocentric interference with the climate system to, inter alia, allow ecosystems to adapt naturally to climate change, to ensure that food production is not threatened*". However, ocean acidification has frequently been perceived more as a threat to climate change rather than an effect of it (Harrould-Kolieb, 2019). This separation of the two phenomena has resulted in placing ocean acidification outside of the mandate of the UNFCCC and contributed to creating a gap in global governance with no multilateral agreements having jurisdiction over mitigation of the increasing of ocean acidification. The strategy of framing ocean acidification as a separate problem to climate change is reflected in its absence from the work of the UNFCCC. Some legal scholars have suggested how to reframe the existing UNFCCC mandate including adopting a new Protocol in order to fill the gap of governance and reframe ocean acidification as an effect of climate change (Kim, 2012).

The other regime relevant for ocean acidification is the Paris Agreement, adopted 23 year later after the UNFCCC Convention, in 2015. The Paris Agreement seeks to strengthen the implementation of the UNFCCC, especially its objective in Article 2. One of the most interesting aspects of the Paris agreement is the attempt to quantify clearly the ambiguous objective of Article 2. According to Article 2(1), the Paris Agreement aims to strengthen the global responses to the threat of climate change, including "*holding the increase in the global average temperature well below 2 C above pre-industrial levels.*" However, it still remains unclear what impact this temperature objective will have on ocean acidification.

Another provision of the Paris agreement relevant to ocean acidification is Article 2.1(a) stating "*…global peaking of greenhouse gas emissions as soon as possible… and rapid reduction thereafter in accordance with best available science, so as to achieve a balance between anthropogenic emissions by sources and removals by sinks of greenhouse gases by the second half of this century*". In that sense some attempts have been made to understand how to respond in law to ocean acidification with mitigation work including the adoption of a separate target for CO^2, the setting of a goal for ocean acidification alongside that of temperature, the possible inclusion of mitigation goals for ocean acidification into the National Determined Contributions (NDCs) and a formal recondition of ocean acidification as a concurrent threat to climate change rather than its effect.

However, there is no threshold of "unacceptable" pH change that could establish a standard to be integrated into treaties and other legal tools, including the UNFCCC. Establishing a threshold or boundary line for ocean acidification that should not be surpassed is much needed. It is still not known what is the exact amount of CO_2 that should be reduced (like emission reductions pathways) in order to bring the level of ocean acidification to an "acceptable level."

The IPCC Special Report on Ocean and Cryosphere of 2019 does not contain such predictions establishing a direct link between the 1.5°C and ocean acidification. There are indeed direct links between cumulative CO_2 emissions, the level of global warming (and thus remaining carbon budgets), and ocean acidification but not between the 1.5°C and ocean acidification.

According to Article 2(1), the Paris Agreement "aims to strengthen the global response to the threat of climate change," including by holding the increase in the global average to limit the temperature increase to 1.5°C above pre industrial levels."

However, the crucial question is: do we know what impact this temperature objective (the 1.5°C) will have on ocean acidification? In other words, do the goals of the temperature targets accurately take into account ocean acidification or pH levels as part of the reduction of risks and impact of climate change? There are indeed direct links between cumulative CO_2 emissions, the level of global warming (and thus remaining carbon budgets), and ocean acidification but there is no direct system of linkage between temperature targets and the reduction of carbon dioxide concentrations in the atmosphere with the pH level in the oceans. There is no certainty that keeping the

[12]Convention on the Prevention of Marine Pollution by Dumping of Wastes and other Matters, 29 December 1972.

[13]Protocol to the Convention on the Prevention of Marine Pollution by Dumping of Waste and other Matters 1996 (as amended in 2006).

[14]The Arctic States are Canada, Denmark (including Greenland and the Faroe Islands), Sweden, Russia, and the United States.

temperature below 1.5°C will guarantee that the level of ocean acidification will be kept at an acceptable level.

The reframing of ocean acidification as a threat of climate change rather than as aspect concurrent with the impact of climate change, as suggested by study of Harrould-Kolieb (2019), would change the focus for action within the UNFCCC as it no longer would be imperative that the Conference of the Parties (COP) acknowledge ocean acidification as being contained within the mandate of UNFCCC to be addressed alongside climate change.

Rather, ocean acidification as an adverse effect of climate change could become an integral aspect in understanding the complete picture of global climate change and the health of the climate system including the threshold's gap. Therefore, it could inform policy choices on emission limits and adaptation strategies. In this way, ocean acidification would no longer be excluded from the work of the Paris Agreement and would no longer would be seen as an additional problem to climate change but rather its attenuation. Furthermore, scientific advances in terms of assessing thresholds for ocean acidification could even become a new parameter by which to measure the success of efforts to combat climate change in terms of risks and impacts.

As to *Biodiversity Conservation Regimes*, the Convention on Biological Diversity (CBD)[15] was established with the three main goals: the conservation of biodiversity, the sustainable use of the components of biodiversity, and the fair and equitable sharing of benefits arising from the commercial and other utilization resources. The importance of the Precautionary principle[16] in the convention is relevant to ocean acidification in the effort to ensure that the absence of scientific certainty is not used as justification for failing to take appropriate measures to safeguard biodiversity. The CBD therefore pays special attention to the problem of ocean acidification given the severe effects that it can have on marine organisms and ecosystems of the planet. The CBD Convention is significant globally because of its Article 8 regarding the establishment of the management of protected areas which concerning the Polar Regions is important for the enhancement of protected areas. In the Arctic, there is also scope for a closer assessment of the development of a more extensive system of marine protected areas.

As to other developments for example within the typology of the *atmospheric pollution global regimes*, the 1999 Protocol to Abate Acidification, Eutrophication and Ground-level Ozone[17] sets ceilings and national emissions standards for four pollutants (sulfur, nitrous oxides, volatile organic compounds (VOCs) and ammonia). Art. 2 of the Protocol states that a central objective is to ensure that a critical load of acidity is not exceeded, including in marine environments. This provides an important example of "acidity threshold" in an environmental treaty, and could represent a source of inspiration for the new standards and thresholds to be incorporated in future laws and decision-making treaty law processes.

Regional Regimes

In the Arctic, ocean acidification is dealt under the umbrella of the Arctic Council (AC) the central instrument of scientific cooperation in the Arctic Region that is soft law in nature. The main role of the AC (Arctic Council) is to protect the Arctic environment. In the AC, there are four core working groups ensuring that the Arctic development takes place responsibly in respect of the environment[18]. One of the working groups deals with marine environment and livelihood: the *Arctic Monitoring and Assessment Programme (AMAP)* also treats ocean acidification. The AMAP monitors and measures levels of anthropogenic pollutants with the purpose of assessing their effects on the Arctic environment by establishing an Arctic Monitoring Assessment Task Forces, the Secretariat of which is based in Norway.

The AMAP is a scientific organism that mainly elaborate reports in details on the status of Arctic ecosystems and identifies the main causes of change by evaluating the impacts and effects of climate change in the Arctic Ocean not only in the fauna and flora but also on the local population.

According to the final results and evaluations, the AMAP prepares a series of recommendations directed to the Arctic States in order to reduce the risks on ecosystems. The AMAP anticipates that ocean acidification, particularly if coupled with ocean warming and deoxygenation, will drive changes in marine ecosystems and impacts on Arctic biota. According to the latest AMAP report of 2018, these changes pose risks to commercial, subsistence and recreational fisheries, as well as to the provision of other ecosystems services in the region[19]. The AMAP couples anthropogenic ocean acidification to the component of pH reduction caused by human activity.

Another piece of regional soft law related of Arctic Ocean acidification is the Kiruna Declaration[20] adopted under the Swedish Chairmanship at the AC. In the Kiruna Declaration, Arctic Ocean acidification was taken into consideration together with other significant Arctic Scientific Studies including Arctic Biodiversity Assessment, Arctic Ocean Review, and the Agreement on Marine Oil Pollution and Preparedness and Response in the Arctic[21]. In the Kiruna Declaration, Arctic Ocean acidification required the AC to continue to take action on mitigation and adaptation and to monitor and assess the state of Arctic Ocean acidification.

An important regional regime applying to Arctic Ocean Acidification is the 1992 Convention for the Protection of the Marine Environment of the North East Atlantic (OSPAR

[15] Convention on Biological Diversity of 5 June 1992.

[16] The Precautionary principle aims to provide guidance in the development and application of international environmental law where there is scientific evidence of *uncertainty*. See de Sadeleer (2010).

[17] Protocol to Abate Acidification, Eutrophication and Ground level Ozone, 20 November 1999.

[18] The four working groups of the AC are: the Arctic Monitoring and Assessment Programme (AMAP), the Conservation of Arctic Flora and Fauna (CAFF), the Protection of the Arctic Marine Environment (PAME) and the Emergency Prevention and Response (EPPR).

[19] AMAP. (2018). 49.

[20] Kiruna Declaration, on the Eight Ministerial Meetings of the AC, MM08—15 May 2013—Kiruna, Sweden.

[21] Agreement on Cooperation on Marine Oil Pollution, Preparedness and Response in the Arctic of 15 May 2013.

Convention)[22]. The OSPAR Convention is the main regime between 15 States of the western coasts and the catchment of Europe, which together with the European Union (EU), cooperate to protect the marine environment of the North- East Atlantic. The OSPAR regime aims at identifying environmental threats and organizing programmes and measures to combat environmental threats effectively.

In the OSPAR Convention, the link between prevention and precaution is ensured by Art. 2 (2) imposing wide-ranging obligations on States parties to "*take all possible steps to prevent and eliminate pollution*". Pollution is defined broadly as in the UNCLOS Convention as "*the introduction by man, directly or indirectly, of substances or energy into maritime area which results, or is likely to result, in hazards to human health, harm to living resources and marine ecosystems, damage, to amenities or interference with other legitimate uses of the sea*". Ocean acidification falls directly into the above-cited OSPAR definition of pollution because it is a process caused by the indirect introduction by humankind of CO_2 into the ocean and it is likely to result in environmental damage to marine ecosystems.

The purpose of the OSPAR convention is thus to protect the ecosystems from the threat of pollution and problems that could jeopardize habitat health, which is particularly sensitive in the Arctic Ocean. For more than 30 years, this instrument has been able to significantly reduce radioactive waste, phosphor, and heavy metals, regulate offshore activities, and provide a precise evaluation of the status of the water's health.

The regulation of Arctic Ocean acidification is ensured by Art. 2 (2) with the relevance of the Precautionary principle that links prevention and precaution where preventative measures are to be taken when there are "reasonable grounds for concerns… even whenever there is not conclusive evidence of a causal relationship between the inputs and the effects". Climate change is one of seven designated work areas of the OSPAR Commission and it is worth noticing that in 2006 the OSPAR Commission published a report on ocean acidification that included a detailed consideration of its marine environmental impacts. The OSPAR Commission also adopted a decision to prohibit placement of CO_2 on or above the seabed.

At the EU level, several instruments have been developed, both of primary or secondary legislation aiming to protect the marine environment[23]. The preferred instruments adopted by the EU legislator to combat ocean acidification are the directives that require member states to transpose these measure at national level to make them directly applicable. For example, Sweden which is both an Arctic State and a EU member State, has implemented a number of EU directives by adopting legal acts at national level such as the Air Quality Ordinance and the Environmental Code. However, not all the EU member states have implemented laws on ocean acidification to this level.

There are actually no studies from a bottom-up approach showing how EU member states have mitigated ocean acidification. Arctic States member of the European Economic Area (EEA)[24] zone do not document a strong research and legislative framework to combat ocean acidification in spite of the fact that ocean acidification observed across the Arctic Ocean has become increasingly apparent, which indicates that more work is needed in this area with regards to ocean acidification (Galdies et al., 2020).

EU Member States action on ocean acidification shows that the current state of European national policies and legislation addressing the ocean acidification problems is, where existent, uncoordinated. Although the ocean acidification problem is acknowledged at higher levels of governance, such as for example at the European Commission level, it is greatly diluted at EU member states level (Galdies et al., 2020). Even though the EU has adopted a number of multilateral agreements, they only address a fraction of ocean acidification. More effective and stronger legal responses to curb ocean acidification problems, including the establishment of an acceptable threshold at EU level to be coordinated with a global ocean acidification threshold are needed.

The EU attempt has been more directed at regulating climate change in general rather than focusing on ocean acidification. The recent European Green Deal[25] has not dedicated special attention to ocean acidification. However, in the field of research, an important effort has been conducted by a recent Horizon 2020 research, the INTAROS, targeting the Arctic Ocean[26].

Overall, rather than focusing on ocean acidification, all the existing previously mentioned regimes only address fractions of the problems and are solely concerned with measures aimed at mitigating and lowering CO_2 emissions and climate change. Even if there were a system of multi-regulatory governance applicable to ocean acidification in general and specifically also applying to the Arctic Ocean acidification context, it seems that there is a lack of coordination between the different regimes.

Ocean acidification has a global nature that requires cooperation among states at all the scales and layers of governance to address it. The law is without any doubt a promising instrument to respond to ocean acidification because it aims at changing the behavior of states and industries, and individuals, but more cooperation between the different layers of governance and actors at all the levels of governance is

[22]Convention for the Protection of the Marine Environment of the North-East Atlantic, 22 September 1992, in force 25 March 1998 ("OSPAR Convention").

[23]A number of Directives are applicable to Ocean acidification and also to the Arctic Ocean, such as for example Directive (EU) 2000/60/EC Water Framework; Directive (EU) 2003/87/EC Emission Trading Scheme; Directive (EU) 2008/50/EC Ambient Air Quality; Directive (EU) 2008/56/EC Marine Strategy Framework; Directive (EU) 2008/Reduction of national emissions of certain atmospheric pollutants.; Directive (EU) 2009/28/EC Measures for promotion of energy from renewable sources; Directive (EU) 2012/27/EC Energy Efficiency; Directive (EU) 2014/89/EU Marine Spatial Planning.

[24]The European Economic Area (EEA) was established via the Agreement on the European Economic Area, an international agreement which enables the extension of the EU's single market to member states of the European Free Trade Association (EFTA). The EEA links the EU member states and three EFTA states (Iceland, Liechtenstein, and Norway) into an internal market governed by the same basic rules.

[25]European Green Deal" (Communication). COM. (2019). 11 December. 640 final.

[26]The overall objective of INTAROS is to develop an integrated Arctic Observation System (iAOS) by extending, improving and unifying existing systems in the different regions of the Arctic. See more at: Available online at: https://cordis.europa.eu/project/id/727890 (accessed May 5, 2021).

necessary. Multiple instruments are applicable to remedy ocean acidification in general but they are not designed to address ocean acidification specifically and to take into account the diversity of ocean acidification impacts in different geographic areas, such as the Arctic Ocean, presents. Even when some aspects are taken into account in some pieces of legislation at global level that are applicable to the regional level and to the Arctic Ocean, there is no coordination between the regional (Arctic Ocean) and the global regimes when it comes to regulating ocean acidification and no coordination between the global regimes and the regional regimes.

ETHICAL AND CLIMATE JUSTICE ARGUMENTS AT THE CORE: THE CRITICAL ENVIRONMENTAL THEORIES CONNECTED TO ENVIRONMENTAL LAW

This section explores how ethical and justice arguments shared by the society are interconnected with environmental law and can shape legal and policy responses to globally deteriorating climate conditions of ocean acidification. The practice from which an unequal distribution of global pollution effects emerges such as like the depletion of common global resources (i.e., the air or the oceans), is unjust.

According to Rawls (2017), in its theory of the history of justice, *inequalities* perpetuated by environmentally destructive practices should not be allowed to continue, as there is reason to believe that these practices do not benefit all parties. Among these practices, resource exploitations do not benefit all peoples but rather undermine the quality of life of a global majority. An appropriate climate ethic takes into consideration *collective action problems* affecting the Arctic, like global warming and increasing ocean acidification. All this environmental degradation is taking place by human hands acting as isolated individuals, business companies, or states.

In the Arctic, resource access and competition for oil and gas, fishing or mining are the main causes of CO_2 emissions, and also reverberate at planetary level, as these resources exploitations are the central cause of global warming. Climate change in the Arctic is one of the greatest threats to the fragile Arctic marine environment causing ocean acidification and melting of sea-ice. Ethical and climate justice arguments should guide how ocean acidification can be talked about and perceived from a legal justice perspective and how such a perspective regarding ocean acidification could alleviate global warming.

There are three main kinds of ethical responsibilities that could arise as a consequence of climate change (Caney, 2009). These ethical responsibilities are: (1) a responsibility to mitigate climate change; (2) to attribute responsibility to enable those hit hardest by climate change to adapt (developing countries), and (3) once liability established, to compensate those affected by the threat. In all these possibilities, the common denominator is how to distribute the burden of responsibility and decide if this responsibility should be distributed among official actors (state and international organizations) or non-official actors (such as corporations, individuals, NGOs, groups of interests, or lobbies) or shared by all at once since climate change is a global, collective problem, such as ocean acidification. Also, the capacity of oceans and marine ecosystems to adapt and function under pH levels of acidification can be framed as a *global collective action* problem. This kind of *global collective action* problem can be understood as the "Tragedy of Commons" (Hardin, 1968). In the same way, oceans and the climatic atmosphere are not inexhaustible. The amount of clean air on earth is, for example, not without limit. The Tragedy of Commons predicts a gradual overexploitation of common pool resources, including oceans and atmospheric resources, which include an unlimited decrease of the pH levels of ocean acidification with deteriorating, environmental and human damaging consequences.

At the core of critical environmental theories applicable to ocean acidification is the discourse regarding co- responsibility for the deteriorating effect on humanity of climate change (such as crop failure, drought, flooding, and ocean acidification). Therefore, ocean acidification is perceived as one of the severe deteriorating effects of climate change.

New standards are directly needed (Skillington, 2017a). One solution is to apply and connect ethical and climate justice criteria. From and environmental law perspective, there are three main factors that can be taken into account to design a model of ethical and climate justice to ocean acidification applicable also more specifically to Arctic ocean acidification which are: to *attribute responsibility*, *minimize uncertainty, establish environmental liability* combined with a *compensation fund approach*

Concerning the *attribution of responsibility*, certain societies have polluted with cumulative emissions for long a period by conducting certain activities, that caused damage to the atmosphere and to the oceans, determining ocean acidification (often defined as "past historical pollution"). These damages includes not only damages to the environment but also to communities (such as for example to indigenous people or communities depending on natural resources and ecosystem managements at sea). According to principles of environmental law, norms and values, the wrongdoers (i.e., states, or international organizations polluting and increasing ocean acidification) should assume responsibility with the consequential compensation to harm others as a consequence of a wrongful act such as for example the violation of primary rules (i.e., treaties) and admit circumstances excluding responsibility for wrongful acts. There are however, a lot of uncertainties on how far is it possible to consider the historical damaging period and which sectors (transport, energy, heating, or food) contributing to the increase of ocean acidification. Another difficulty is due to the fact that the extent to which different activities contributed to acidification vary substantially between different geographical localities, and to identify responsibility in a "global commons" areas[27] may be a hard task.

[27] "Global commons" in international law are areas that do not fall within the jurisdiction of any one country and are defined as "international or global commons." The notion of global commons posits that there are limits to national sovereignty in certain parts of the world and that these are areas should be open to use by the international community but closed to exclusive appropriation by treaty

Therefore, *minimizing uncertainties* becomes relevant because increase the possibility to activate responsibility. For example, there is *uncertainty* as to which actors may also affect the impacts of ocean acidification via interacting stressors because ocean acidification could be aggravated also by other stressors such as eutrophication, wastewater discharge, or fishing and some other actors may also contribute to ocean acidification at multiple scales to make marine ecosystem more vulnerable to the impacts of ocean acidification.

Environmental liability is just one aspect of responsibility and arises out from activities not prohibited by international law that cause damage. The aim of liability is the prevention of environmental damage and reparation of victims instead of stopping the activity and has more a preventative function. The relevant actors are states or international organizations. However, there is still the need to understand what is the clear terminological distinction between the terms "responsibility" and "liability."

Establishing liability as a consequence of environmental damage is also related with the problem that the environment (in this case the sea or ocean) which is a public goods or *res propriare*, or *res communes*, or *res nullus* is not belonging to anyone. Liability requires the existence of "standards" of justice. Violations of these standards may be subject to legal investigation, which differs according to the *common law* systems (based on tort law) and *civil law* systems. There are three main types of civil liability criteria as a consequence of environmental damage: (1) fault based, (2) strict liability, and (3) absolute liability (Cassotta, 2012). In this article, only the two types of liability will be considered as relevant for the purpose of this section. The typical situation where *fault liability* apply is when subject × damage subject y with the existence of the subjective element of *culpa*, and x must repair. However, in environmental law civil liability when applied to environmental damage is completely disrupting this situation because the good belonging to subject y (in our case the environment that has been damaged) is not susceptible to be object of ownership as it is a public good and does not belong to anyone. Under a *strict liability* regime the victims of the environmental damage are facilitated since they do not have to prove *culpa* of the potential wrongdoer in order to receive compensation. *Strict liability* is generally adopted in case of involving activities likely to have harmful consequences even if conducted with due care or due diligence or in respect of the normal criteria for standards or tolerability. In case of *fault liability*, the injured parties (y) of the environmental damage are not facilitated since they must prove the *culpa* of the potential wrongdoer (x) in order to be compensated for damages. A certain development in terms of environmental justice is when civil liability as a consequence of environmental damage is connected to the Polluter-pays principle.

According to Kramer's definition, the Polluter-pays principle is "firstly an economic principle belonging to the public sphere, and has to be understood as expressing the costs of environmental impairment, damage and clean-up that should not be borne via society's taxes, but by those persons who caused pollution" (Krämer, 2007).

However, there are several *uncertainties* concerning the relationship between the Polluter-pays principle and the sphere of civil liability such as: who the polluter is, what is the environmental damage and how much compensation should be paid. Environmental law still has to evolve in order to make sovereign states liable for environmental damage, ocean acidification included.

There are still issues in relation to *causation* or *probabilistic causation* relevant to climate change such as the nexus between the author of the damage and the event which is difficult to prove and *cumulative emissions* or *diffuse pollution* which is the problem emerging when the damage is not a consequence of a single damaging situation. In the latter case, the difficulty is to determine the source of pollution and the percentage of responsibility of each polluter, and establish if there was joint, several responsibility, and the percentage of responsibility attributable to each. Recent rulings have progressed with the ordering to the Netherlands to reduce GHGs emissions by 2020 and there is progress with climate modeling to better determine and identify the source of pollution. However, causation is still a non-linear and challenging problem. The ethical and climate justice problem is huge since cumulative climate harms is generated by many states and industrial actors, all engaging in pollution practices, all increasing ocean acidification.

The Courts of Hague attempt to hold states accountable for not meeting CO_2 emissions targets, as the general understanding is that high polluting states are *jointly and several responsible* for global warming and ocean acidification of the oceans as well as the reduction of subsurface oxygen levels affecting the growth of marine phytoplankton, coral reefs and fish stocks (Skillington, 2017a). Agents inflicting the increase of ocean acidification has not to be attributed to states only but also to corporate actors, especially industries in the sector of fossil fuels, which are emitters of GHGs emissions.

The idea of preventing or moderating negative consequences of ocean acidification suggests that someone takes responsibility for limiting actions of individuals that increase common risks. This could be done, for example, by transferring the uncompensated damages of ocean acidification to the top fossil fuel companies. This perspective suggests that it is the fossil fuel companies that increase climate changes impacts determining ocean acidification. However, governments deal with risks and externalities as well[28]. In the face of ocean acidification, it is unlikely that only the public sectors and governments are solely responsible but also the private sector which often has no incentives to change its "business as usual" practices. It would not be equitable and just to expect governments to be the primarily responsible parties and financiers of negative effects of ocean acidification. An innovative solution in that sense could be to transfer the financial risks of ocean acidification to fossil fuel

or customs. Examples of "global commons" areas are: the High Seas, Antarctica, Outer Space and the Atmosphere. See Redder and Hughes (2008).

[28]The concept of "externalities" refers to the activity of the potential polluter. The potential polluter is in this way forced to also include in its costs for production, the costs that could emerge from environmental damage through a mechanism called "internalization." Cropper and Oates (1992).

companies. The proposal in this line of reasoning could be to establish an ocean acidification fund as has been done in the past to clean up and compensate victims for a number of hazardous activities such as for example, for oil pollution spills, toxic chemical and asbestos contaminants and opt for a "compensation fund for ocean acidification."

The *compensatory fund approach* is based on precedents in law where it is not possible to identify the author of the damage to the environment and thus funds are created where upstream taxes, levies or excises are imposed on the introduction of harmful substances. Important examples of compensatory fund approach are: (1) US Superfund scheme and CERCLA (Comprehensive Environmental Response, Compensation and Liability Act of 1980)[29]; (2) three international conventions on oil pollution which are the 1992 International Convention on Civil Liability for Oil Pollution Damage [International Oil Pollution Compensation Fund (IOPC Fund), 1992]; the 1992 International Convention on Civil Liability for Oil Pollution Damage (1992 CLC Fund)[30]; and the 2003 International Supplementary Fund for Compensation for Oil Pollution Damage[31], and (3) the US oil pollution regime called 1990 US Oil Pollution Act (OPA)[32] All these regimes have in common that they be based on precedents for upstream levies, taxes and excises on feedstock.

CERCLA establishes a "Superfund" financed primarily by excise taxes on petroleum and chemical feedstock, to enable governments to pay for the clean-ups of hazardous chemicals. The reason why CERCLA is extremely interesting for a possible applicability in a legal framework for loss and damage caused by ocean acidification is precisely that it can compensate even if the polluters are not identifiable at the origin. The aim of this fund is to compensate from damage deriving from atmospheric pollution to the subject or parties that cannot find any solutions from the civil liability mechanism. This is exactly the case of CERCLA. In CERCLA there is no need to demonstrate causation, as what is relevant is to identify the cause of harm and attributes strict liability. Compensation funds can be combined with strict liability as it occurs in numerous international conventions, as all those identified in the previous paragraphs.

Other examples are those of the triadic regime previously identified the (1) 1992 CLC Fund, the (2) 1992 IOCP Fund, and the (3) the 2003 International Supplementary Fund for Compensation for Oil Pollution Damage). Under the international law of treaties, international conventions protecting the environment contain compensation schemes on transboundary pollution. In order to design a regulatory framework for loss and damage applicable to ocean acidification, some elements of each can be picked up. The relevance of these conventions with respect to ocean acidification is especially due to the possibility to apply the Polluter-Pays principle. In general, liability and state liability rules determine whether the Polluter-pays principle really applies or if it is just a "principle in the air" and the regimes on oil pollution can fill this gap because these regimes are examples of "canalization"[33] of liability.

In particular, the triadic regime provides three layers of compensation available for victims of pollution damage. The first layer of compensation derives from the 1992 CLC Fund, which covers damage caused by oils of a state party of the convention and where both joint and several strict liability are placed on the owner of the ship from which the pollution escaped. The second layer of compensation arises with the 1992 IOCP Fund providing supplementary funding, where compensation available under CLC is insufficient. As in CERCLA, compensation payments are financed by the contributors; private companies or other entities (private or public). The third layer is the 2003 International Supplementary Fund for Compensation for Oil Pollution Damage, which is also financed by contributors. The 1990 US OPA established the Oil Spill Liability Trust Fund which like CERCLA allows trustees to spend up US 2.5 billion for removal costs and damages for each incident and the interesting aspect in connection to ocean acidification is that it provides parties to be liable also for the diminution in the value of natural resources and not only on the costs of restoration. However, environmental law factors need also to be connected to ethical and climate justice criteria in order to design a model of ethical and climate justice to ocean acidification and establish new standards.

The design of a proposed fund, as discussed previously can overcome climate ocean injustice created by ocean acidification, overcome the weakness of civil liability to compensate for ocean acidification. In a more concrete way, by combining CERLA to the global and US oil pollution regimes, the fund should include issues of: (1) identification of liable parties, which are the companies that should be levied for their annual historical production of fossil fuel, (2) identification of the claimants which are the victims vulnerable to ocean acidification that can bring claims against the fund by a state party claiming on behalf of its affected citizens, (3) establishment of a levy without limitation, like in the case of the 1992 IOPC Fund and CERLA, and (4) provide parties liable also for the diminution in the value of natural resources and not only on the costs of restoration, as in the case of the OPA, even if the quantification of the damage cause by ocean acidification could be difficult to calculate.

Existing theories of climate change justice mainly focus on corrective and distributive justice approaches aiming at attributing responsibility to developed countries to take the lead on climate change (Lyster, 2015). The main theories of climate change justice that are applicable to ocean acidification are classified in three types: (1) Contribution to the problem as corrective approach based on the Polluter-pays principle (2) ability to pay principle as distributive justice and a (3) hybrid

[29] 42, U.S.C. ss. 9601 et seq., as amended through P.L. 107-377, 31 December 2002, see also at https://www.epa.gov/superfund/superfund-cercla-overview.

[30] London (UK), 27 November 1992, in force 30 May 1996, see at https://www.imo.org/en/About/Conventions/Pages/International-Convention-on-Civil-Liability-for-Oil-Pollution-Damage-(CLC).aspx.

[31] See at https://iopcfunds.org/

[32] Available online at: https://www.epa.gov/enforcement/oil-pollution-act-opa-and-federal-facilities

[33] "Canalization" means to "canalize" or "channel liability" toward the person who is in control of the activity "ex ante." In case of oil pollution and pollution at the sea, the person who is in control of the ship or the ship owner or in case of nuclear pollution, it will be the operator of the nuclear power plant.

approach merging corrective approach and distributive justice (Lyster, 2015).

These theories can be interconnected with environmental law principles, and concepts. They can be applied specifically to Arctic Ocean acidification for an understanding of how climate justice for the Arctic Marine Environment could shape the behavior of actors such as states, business sectors, industries or individual and how actors should think and act differently.

Contribution to the Problem—A Corrective Approach

The contribution to the problem in terms of corrective justice is how to connect civil liability to the Polluter-pays principle to minimize the *uncertainties* caused by unsolved enigmas such as the identification of the author of pollution, quantification of the damage, compensation and problem of a time factor, or the so-called remoteness of the damage or historical pollution or emissions.

The obligation to pay for climate change (and the impacts on ocean acidification) should be bridged with both the capacity to pay in terms of income and responsibility to pay (historical emissions) in the interests of equity (Skillington, 2017a). A corrective justice approach is based on the idea that it is the countries that mostly caused pollution in terms of global cumulative emissions that should contribute to the costs. Ethical justice in the corrective approach is based on the concept that developed countries have the ethical responsibility to reduce their emissions given their cumulative, diffuse and "past historical pollution" and correct the negative impacts determining ocean acidification. The determination of responsibility for climate harm is based on evidence generated by the IPCC on existing patterns of GHGs emissions among states.

Ability to Pay Principle—A Distributive Justice

The ability to pay principle in the idea of distributive justice that derives from the claim that developed countries have a greater capacity to cover the costs of mitigation and adaptation to climate change. The distribute justice approach differs from the corrective approach because it does not focus on who contributes to the problem and who is responsible but rather on who has the capacity to rectify the harm and who can mitigate the problem. What is a need is to establish a fair system of climate change mitigation that tackle the different levels of inequalities existing in individual countries. The assessment of capacity to pay links state's responsibility with individual with responsibility leading to the possibility to change individual behavior and shape actions. Distribute justice does focus on weighing and calculating the capacity to pay and responsibility to pay in terms of the distribution of income and emissions across populations within each state. The ability to pay and distributive justice have the potential to enhance a type of responsibility that changes consumption behaviors, habits, and engages the awareness to conserve the environment and pay the price for not polluting, including the price of for not increasing ocean acidification.

A Hybrid Approach—A Corrective and Distributive Approach Leading to Legal Cosmopolitanism

The hybrid approach combines and integrates the corrective approach with the distributive justice approach by integrating "who contributes to the problem" with "who should pay." The approach takes into consideration different problems related to negative externalities, causality link, and *uncertainties*. Property rights at seas are often poorly defined especially outside the Exclusive Economic Zone (EEZ) and emissions of CO_2 have impacts far away from their sources. National policies could provide incentives to internalize the costs. In order to account for the true costs of carbon on global warming, these must also be globally coordinated, address all the sources of carbon and provide for an account for climate change including changes due to CO_2 and their ocean acidification effects (Turley and Gattuso, 2012).

However, true costs of carbon on global warming are not only uncoordinated they do not even cover all the emitting industries and do not account for ocean acidification (Sterner and Coria, 2019). This approach also requires that the financial benefits agreed to in multilateral negotiations reach individuals and are not simply distributed to states. This approach has, therefore "cosmopolitan implications" deriving from the cosmopolitan justice view of justice being a global responsibility rather than state-based. Legal cosmopolitanism is concerned with the legal status of individuals as human beings, rather than citizens of specific states.

Cosmopolitan heritage is debated in light of a range of pressing concerns, including persisting inequalities, the rapid loss of biodiversity, the depletion of resources including ocean acidification, the loss of home, livelihood and natural habitat, unmitigated climate change that impact adversely on the region of the Arctic (Skillington, 2017b). Kant's political theory bases cosmopolitan law on the need to protect the rights and dignity of all individuals (Kant, 1975). This view could instill a new idea to establish an obligation on fossil fuel producers to contribute to a Fund of climate ocean acidification disaster response, which might also be regarded as "cosmopolitan."

There is an undeniable link attributing ocean acidification to the "supply side" of fossil fuels production and global extraction projects causing anthropogenic CO_2 emissions. Over the past 200 years, the world's oceans have absorbed more than 15 billion metric tons of carbon dioxide emitted from human activities.

The atmospheric concentration of CO_2 has increased because of the burning of fossil fuel such as coal, gas, and oil along with land use change such as the conversion of natural forest into crop production. Continuing with this "business- as-usual" scenarios, in the end of the century the surface waters of the ocean could be nearly 150 percent more acidic, resulting in a pH that the oceans haven't experienced for more than 20 million years.

The polluting exploitation behavior typical of the current Anthropocene Era has contributed to the degradation of the oceans and the marine environment. Climate responsibilities extend well-beyond official actors, such as states, national governments, International Organizations or institutions.

Also non-official actors such as sub-national governments, corporations, utilities and individuals are polluting agents and liable for ocean acidification contributing damage not only to the marine environment but also to local communities. In societies that are completely interlinked to the marine environment and which are strongly dependent on marine biodiversity, ocean acidification alters not just the chemistry of oceans but also their livelihood. Many indigenous and coastal communities, such as those across the Arctic, for example, have felt the effects on the food-web in the Arctic Ocean which is very sensitive.

Therefore, a significant increase of the population of one species or the disappearance of another could have dramatic damaging and loss effects on the entire Arctic marine ecosystem and on indigenous people which are part of the ecosystem. Non-official actors are therefore viewed as having an obligation to address climate change and decrease ocean acidification.

Societal perceptions that fossil fuels companies bear distinctive climate responsibilities are reflected in the existence of movements of reaction and climate lawsuits against plans of exploitations pushing for a shift in behavior toward a transition from fossil fuel to non-polluting green activities in line with the goals of the Paris Agreement.

Climate change litigation is one example of legal actions against fossil fuel companies' plans to exploit and the governments allowing exploitations licenses for fossil fuel's extractions. One example, in the Arctic is with the issuance in 2016 by the Norwegian Ministry of Petroleum and Energy of an offer of 10 new production licenses in the 23rd licensing round on the Norwegian continental shelf under the Barents Sea (Stokke, 2020). Five months later three Norwegian environmental organizations filed a suit against the government. In 2017 and 2018, for example more than a dozen US cities and counties and the state of Rhode Island filed suits against several investor-owned fossil fuel companies seeking to hold them liable for their contribution to the damages from sea level rise and increasingly extreme weather that climate change is imposing on local communities.

Recent research has quantified the contribution of CO_2 and CH_4 emissions traced to the products of major fossil fuel companies and cement manufacturers to global, atmospheric CO_2 surface temperature and seal level rise. This means that major industrial carbon producers can be linked to responsibility aspects connected to societal considerations as a consequence of ocean acidification. Loss and damage in regions that are affected by ocean acidification in the context of climate change and other stressors is now identifiable which can open the path to future advancements in terms of attributing responsibility to major fossil fuel producers for the current and near-term risks of further loss and damage to human communities dependent on marine ecosystems and fisheries vulnerable to ocean acidification.

However, the extent and severity of future damage as a consequence of ocean acidification and climate change on the marine species and ecosystems, and the human communities dependent upon them will continue to be determined by the future course of plans to exploit natural resources by the fossil fuels industries. For example, in the Arctic Ocean, which is characterized by high fishing potentials, high value fisheries such as those harvesting Alaska red king crab and Atlantic sea scallop, decreases as a consequence of ocean acidification as assessed by projections, eventually become apparent in the next 20 and 30 years when effects exceed natural variations together with other projections.

Specific climate impacts and damages due to ocean acidification are attributed not only to states entities but also to non-official actors mirroring a situation of oceanic atmospheric *injustice* caused by a carbon society that undermines the idea of climate democracy and increases a situation of climate ocean *injustice* that will lead the Arctic Ocean and the Oceans in general to a point of non-return in terms of sustainability, marking the collapse of one of the most relevant planetary boundaries.

The design of a proposed fund, as discussed previously can overcome climate ocean injustice created by ocean acidification, overcome the weakness of civil liability to compensate for ocean acidification. In a more concrete way, by combining CERLA to the global and US oil pollution regimes, the fund should include issues of: (1) identification of liable parties, which are the companies that should be levied for their annual historical production of fossil fuel, (2) identification of the claimants which are the victims vulnerable to ocean acidification that can bring claims against the fund by a state party claiming on behalf of its affected citizens, (3) establishment of a levy without limitation, like in the case of the 1992 IOPC Fund and CERLA, and (4) provide parties liable also for the diminution in the value of natural resources and not only on the costs of restauration, as in the case of the OPA, even if the quantification of the damage cause by ocean acidification could be difficult to calculate.

DISCUSSION AND CONCLUSION

The issue of responsibility concerning ocean acidification is actually profoundly linked to fossil fuel in the Arctic Ocean where we witnesses a real race for natural resource' extraction. The huge oil reserves there attract the oil companies to extract. These companies have plans to extract natural resources like oil and gas until 2030. It is very difficult for governments to make a break with these plans because there are enormous interests at stake. However, if we persist in the extraction of all the resources mankind wants to take the result will unquestionably be increasing ocean acidification. Taking into account science predictions, the temperature will rise 4 degrees Celsius by 2030. Not only states are responsible for ocean acidification but also isolated individuals, companies, corporations acting in the market.

The increasing of ocean acidification due to fossil fuel activities interacting with other activities that are exploiting and depleting the marine environment, contradicts the concept of sustainable development, and the goals of the Paris Agreement. It will contribute to ice melting which will determine Sea Level Rises (SLR) and in turn will have a direct impact outside the Arctic, such as for example on the Pacific Islands and the sinking of these Islands such as for example the Small Island and Developing States (SIDS).

Only by taking ocean acidification into account as a collective action problem and linking it to issues of responsibility by multiple actors as a threatening consequence of climate change, and with an appropriate ethical and climate justice perspective, will it be possible to find solutions. With regards to Arctic ocean acidification, a smart system of energy transition involving all the actors that are responsible for ocean acidification to make them change behavior should be organized at regional level, i.e., at the Arctic Council level. Here a *forum* of discussion, together with a special Fund for compensation financed by those who are the most responsible for ocean acidification, could represent an avenue to avoid the erosion of one of the most relevant planetary boundaries of our Earth system. The fund could be an innovative vehicle for transferring the financial risks of ocean acidification to fossil fuel companies by establishing synergistic linkages between the different levels of multi-regulatory scale, which are now absent. A fossil-fuel Fund for Arctic ocean de-acidification could set the foundation to connect legal, environmental principles to climate justice, which could serve as a test case for a general funding of global de-acidification at a global level.

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Complex Vulnerabilities of the Water and Aquatic Carbon Cycles to Permafrost Thaw

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The spatial distribution and depth of permafrost are changing in response to warming and landscape disturbance across northern Arctic and boreal regions. This alters the infiltration, flow, surface and subsurface distribution, and hydrologic connectivity of inland waters. Such changes in the water cycle consequently alter the source, transport, and biogeochemical cycling of aquatic carbon (C), its role in the production and emission of greenhouse gases, and C delivery to inland waters and the Arctic Ocean. Responses to permafrost thaw across heterogeneous boreal landscapes will be neither spatially uniform nor synchronous, thus giving rise to expressions of *low to medium confidence* in predicting hydrologic and aquatic C response despite *very high confidence* in projections of widespread near-surface permafrost disappearance as described in the 2019 Intergovernmental Panel on Climate Change Special Report on the Ocean and Cryosphere in a Changing Climate: Polar Regions. Here, we describe the state of the science regarding mechanisms and factors that influence aquatic C and hydrologic responses to permafrost thaw. Through synthesis of recent topical field and modeling studies and evaluation of influential landscape characteristics, we present a framework for assessing vulnerabilities of northern permafrost landscapes to specific modes of thaw affecting local to regional hydrology and aquatic C biogeochemistry and transport. Lastly, we discuss scaling challenges relevant to model prediction of these impacts in heterogeneous permafrost landscapes.

Keywords: permafrost, boreal, carbon, hydrology, climate change

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INTRODUCTION

Permafrost, ground remaining below 0°C for more than 2 years, contains roughly a third of the world's soil organic carbon (Schuur et al., 2015; Meredith et al., 2019) and covers nearly one-fourth of the terrestrial Northern Hemisphere (Brown et al., 2002), including much of the boreal forest and tundra biomes (**Figure 1**). The state of permafrost and the corresponding sequestration of frozen soil organic carbon are not permanent. Disproportionally high rates of warming in northern latitudes increase thermal stress on permafrost, particularly warm permafrost near 0°C (James et al., 2013; Meredith et al., 2019). The fate of permafrost soil organic carbon has received attention in the past decade among the scientific community, greenhouse gas policy strategists, and the general public because its decomposition could result in enhanced emissions of carbon dioxide (CO_2) and methane (CH_4) to the atmosphere, promoting a positive feedback to atmospheric warming

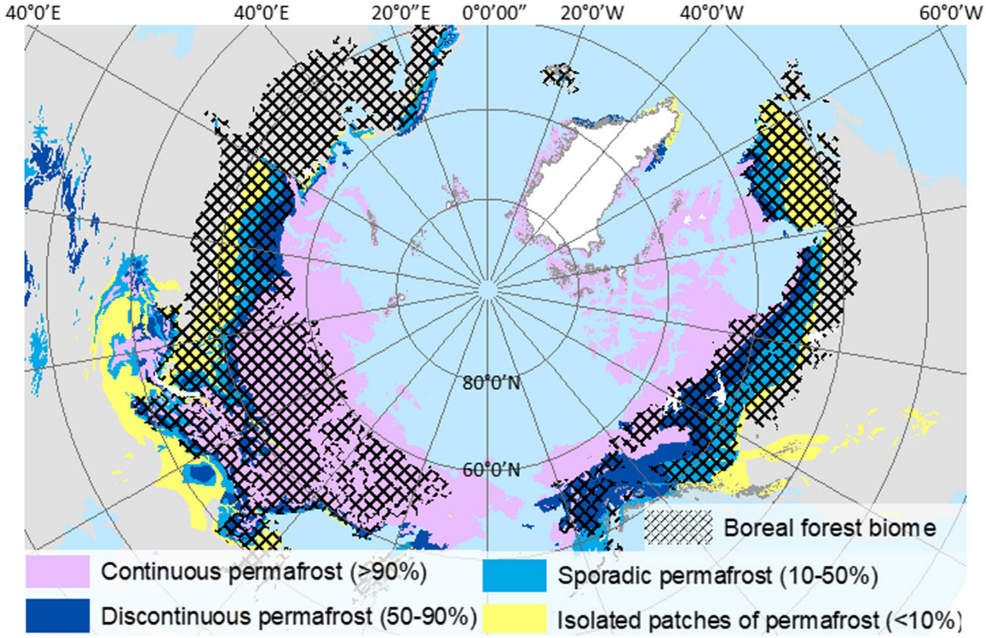

FIGURE 1 | Permafrost coverage in the northern hemisphere (from Brown et al., 2002). Stipled pattern overlay represents the boreal forest biome coverage.

(Schuur et al., 2015). Large-scale model projections of future permafrost C release calculate CO_2 and CH_4 emissions that are spatially variable across landscapes (Grosse et al., 2016; Xia et al., 2017; McGuire et al., 2018). These projections are limited by several sources of uncertainty (i.e., vegetation shifts, biomass accumulation factors, organic carbon (OC) degradability, N availability, subsidence, and permafrost thaw rates) (Lawrence et al., 2015; Abbott et al., 2016; Schädel et al., 2018) with one of the largest being the response of the lateral hydrologic-C flux that is generally excluded or treated simplistically in large-scale models (Ala-aho et al., 2018; Hugelius et al., 2020; Tank et al., 2020). Although C export from some Arctic permafrost systems is frequently conceptualized as one-dimensional CO_2 and CH_4 emissions to the atmosphere, permafrost regions are also influenced by C lateral transport including above-ground pathways (e.g., streams) and below-ground pathways (e.g., groundwater) that transport particulate and dissolved inorganic and organic C (Stackpoole et al., 2017; Plaza et al., 2019; Wild et al., 2019). Potential ecosystem effects stimulated by lateral C transport include the release of colored terrestrial dissolved organic matter to the ocean, altering marine primary productivity (Spencer et al., 2009) and the release of stored contaminants associated with permafrost OC, such as mercury, to food webs (Schaefer et al., 2020; Zolkos et al., 2020a). The added complexity of lateral C transport requires comprehensive analyses that integrate hydrologic and biogeochemical considerations (Vonk et al., 2019; Tank et al., 2020). Unlike the atmospheric release pathway of permafrost C, in which high source permafrost C content generally corresponds to high potential C-gas emission, lateral hydrologic C transport potential is jointly influenced by C source and thaw-impacted surface and subsurface hydrologic flowpaths, which are largely unknown. Tank et al. (2020) describe a state factor approach for exploring aquatic biogeochemical responses to permafrost thaw evaluated through regional case studies across the Arctic. They note that challenges in implementing this approach arise in regions with spatial heterogeneity in state factors, particularly in discontinuous permafrost. We posit this is due to the complexities associated with changing surface and subsurface flowpaths, fluxes, and soil-water residence times in transitional continuous to discontinuous permafrost landscapes. This overview explores these complexities through synthesis of recent work in northern latitudes, particularly at the continuous (>90% coverage) to discontinuous (50–90% coverage) permafrost transition, where shifts from predominately shallow, warm-season flow processes toward deeper year-round flow processes are expected to impact hydrologic C lateral transport. Based on this synthesis, we offer a framework for assessing vulnerabilities of water and aquatic C cycles in unstudied Arctic-boreal areas by addressing landscape, subsurface hydrologic, and C source factors that influence coupled hydrologic and biogeochemical responses to five distinct modes of permafrost thaw. Vulnerability assessments that only account for C source and thaw rate miss the potential effects of hydrologic C lateral transport, a large uncertainty of growing relevance in thawing discontinuous permafrost landscapes (Vonk et al., 2019). Science gaps and research opportunities for advancing vulnerability assessments are highlighted.

FACTORS INFLUENCING VULNERABILITY TO PERMAFROST THAW

Large-scale losses in permafrost extent are expected by the end of this century (e.g., Zhang et al., 2014; Pastick et al., 2015; McGuire et al., 2018; Parazoo et al., 2018; Meredith et al., 2019). Although there are uncertainties in estimated rates and spatial patterns of permafrost loss, these losses are projected to have a major influence on the strength of boreal regions as net C sinks or sources (Grosse et al., 2011, 2016; Mishra et al., 2013; McGuire et al., 2018; Parazoo et al., 2018; Hugelius et al., 2020), and are highly likely to contribute to climate-related economic damages to infrastructure (Melvin et al., 2017). Ecosystem responses will depend on the mode of permafrost thaw and its associated physical, hydrologic, and biogeochemical impacts. It is therefore critical to distinguish among primary modes of permafrost thaw and identify landscape characteristics that influence susceptibility to each thaw mode.

Modes of Permafrost Thaw and Constraints on Detection

Active layer thickening is described by thaw occurring at the top of permafrost that effectively increases the thickness of the layer that seasonally oscillates in temperature above and below 0°C (**Figure 2A**). Active layer thickening tends to be observed as a gradual (press) response to northern latitude atmospheric warming trends and/or other near-surface perturbations with reported thaw rates generally not exceeding a few centimeters per year in areas of occurrence (Streletskiy et al., 2015). Repeat point measurements of the active layer over time can be made to determine thickening *via* manual frost probing, temperature monitoring, and geophysical techniques (Briggs et al., 2017; Douglas et al., 2020). Support for uniform active layer thickening is lacking due to sparse historical data and/or complex and potentially counterbalancing interactions among factors controlling near-surface thermal conditions such as soil properties, snow cover, moisture, and vegetation (Park et al., 2013; Clayton et al., 2021). The lack of a universal protocol that accounts for change in ground surface elevation datum in settings where ground subsidence and compaction accompany thaw also contributes to inconsistencies in active layer observations (Rodenhizer et al., 2020). Integrated modeling and analysis frameworks combining field and remote sensing observations, from both airborne radar and global satellite platforms, continue to improve estimates of active layer thickening (e.g., Yi et al., 2018; Chen et al., 2020).

The progression from active layer thickening to supra-permafrost talik formation occurs when the depth to permafrost exceeds the depth of seasonal freezing. This perennially-thawed zone, or talik, typically develops (**Figure 2B**) over timescales of a few decades (press response) and can reach depths as great as several meters (Streletskiy et al., 2015; Connon et al., 2018). Perturbations, such as wildfire and flooding, can accelerate these processes, promoting talik development within a few years post disturbance (Grosse et al., 2011; Brown et al., 2015; Jepsen et al., 2016; Minsley et al., 2016). Taliks may be spatially isolated (e.g., Rey et al., 2020) or may be laterally extensive and hydrologically connective (Gibson et al., 2018; Terry et al., 2020; Ackley et al., 2021), a distinction that is critical in determining their hydrologic consequences. Taliks and their connectivity are difficult to characterize across broad areas because current permafrost mapping efforts and capabilities at large spatial scales tend to be restricted to near-surface conditions that are shallower than the depths of talik formation (O'Neill et al., 2020). Furthermore, shallow permafrost analysis using satellite and airborne remote sensing data typically employ the assumption that the seasonally-thawed active layer lies directly above permafrost with no perennially-thawed layer. Detection of aufeis, or icings, through remotely sensed optical and thermal imagery and their connection to groundwater upwelling provides a promising means of scoping taliks at large scales (Glass et al., 2020). Permafrost thermal modeling suggests that lateral taliks will become increasingly prevalent across Arctic and boreal regions in the coming decades (Parazoo et al., 2018) and serve as a precursor to accelerated (and irreversible) thaw (Devoie et al., 2019; Walvoord et al., 2019).

Complete permafrost loss refers to broad degradation in a terrestrial setting (**Figure 2C**). Open lake/river talik development also involves the elimination of permafrost at depth but is distinguished here as underlying a lake or river (**Figure 2D**). Detection and monitoring of complete permafrost loss and open taliks require deep-seeing subsurface characterization methods to confirm complete thaw and/or quantify rates of loss with time lapse information. The depth of interrogation required for such determination depends on the maximum depth of permafrost and can range from 10^0 to 10^3 m, limiting extensive characterization. Open taliks beneath lake and river corridors surrounded by thick permafrost have been inferred from airborne electromagnetic surveys (Minsley et al., 2012; Rey et al., 2019) and thermal modeling (Rowland et al., 2011; Wellman et al., 2013). Complete permafrost degradation over land or under water bodies tends to occur gradually (press response) over timescales of centuries or greater (McKenzie and Voss, 2013; Wellman et al., 2013) except in very thin, warm permafrost where complete loss can occur more rapidly (Quinton et al., 2011). Rates of open talik development beneath water bodies are typically greater than beneath terrestrial settings; both can be accelerated if heat transfer *via* groundwater flow contributes to thawing.

When ice-rich permafrost thaws, reduction of pore space associated with ice melt can cause deformation of the ground surface *via* notable subsidence and/or erosion, resulting in various forms of thermokarst development (**Figure 2E**) (Jorgenson and Osterkamp, 2005; Jorgenson, 2013). Though minor ground subsidence may occur upon permafrost thaw in areas of low permafrost ice content, high ice content (ice content in excess of thawed porosity) is a key element controlling thermokarst potential (Vonk et al., 2019; Saito et al., 2020). Thaw-driven subsidence in ice-rich permafrost landscapes produces distinct thermokarst features including thaw lakes, drained basins, water tracks, and bogs and gullies in low relief settings. In settings with moderate to high topographic relief, thaw slumps and active-layer detachment slides commonly occur and develop rapidly through a summer season (pulse response). The ability to

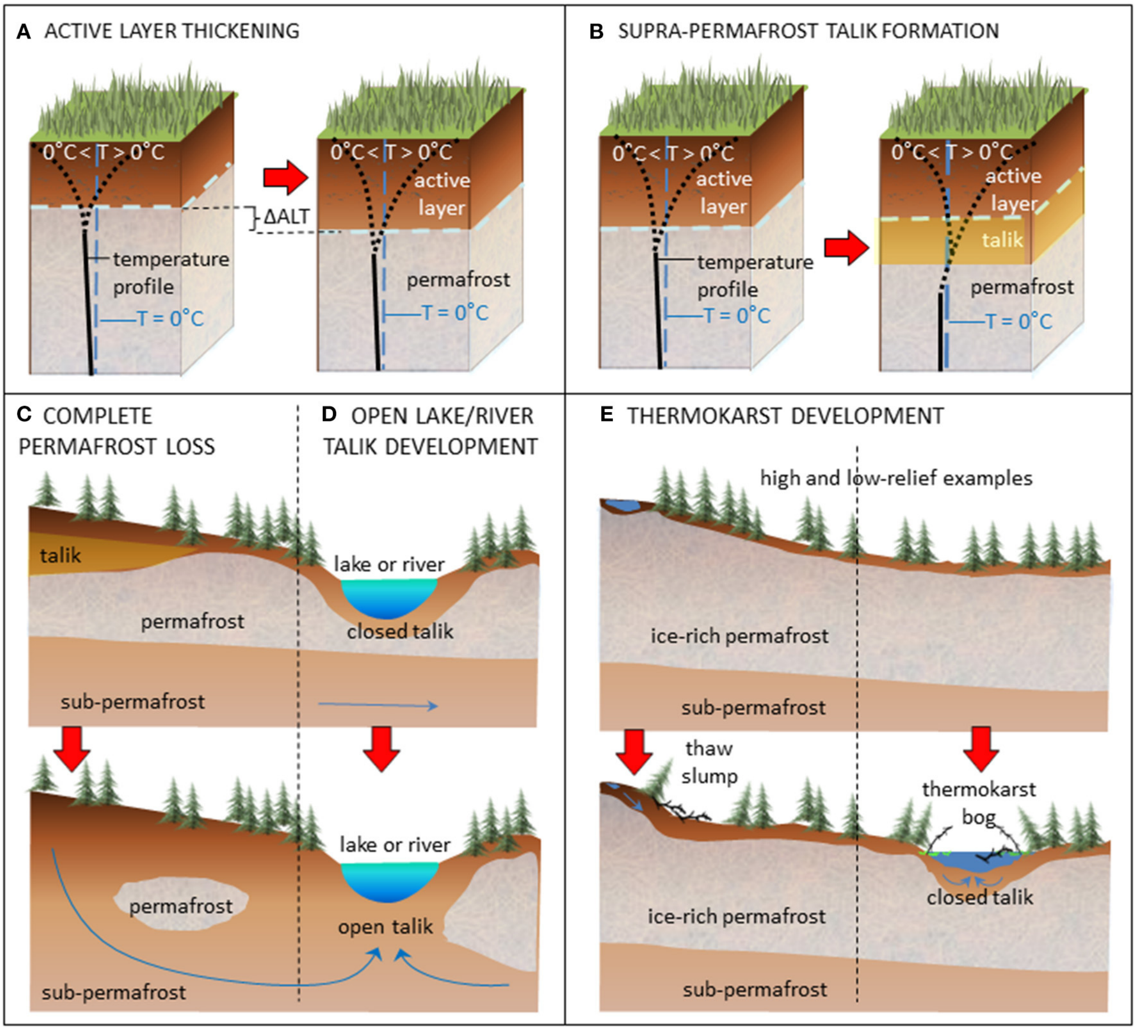

FIGURE 2 | Five primary modes of permafrost thaw **(A–E)** modified from Kurylyk and Walvoord (2021). Red arrows point in the direction of thaw evolution from one state (left or above) to the next (right or below). ΔALT in **(A)** = change in active layer thickness. Blue arrows in **(C–E)** indicate the direction of subsurface water flow.

remotely detect thermokarst features continues to improve with advances in sensor capability, multi-sensor data fusion, and data processing including synthetic aperture radar interferometry (Jorgenson and Grosse, 2016; Nitze et al., 2018, 2020; Abe et al., 2020).

Climatic and Landscape Characteristic Controls on Thaw Susceptibility and Rate

Permafrost distribution is primarily controlled by coupled climatic and terrain factors that influence heat exchange between the atmosphere and the ground surface (Jorgenson et al., 2010). Because paleoclimate and glacial history have a legacy effect, permafrost is not commonly in equilibrium with the current near-surface energy balance, thereby requiring considerations for mapping its distribution, thickness, and depth (Zhang et al., 2014). Large discrepancies of nearly five-fold in estimated current permafrost distribution account for some of the high uncertainties in projected permafrost loss reported by the 2019 IPCC report (Meredith et al., 2019). For instance, 30–99% (69 ± 20%; likely range) permafrost loss is projected under Representative Concentration Pathway (RCP) 8.5. The projected loss range and associated uncertainty is substantially reduced for all RCP scenarios for the subset of models that estimate similar current permafrost distributions. Modeling approaches that consider paleoclimate conditions with extended spin up durations and/or infusion of ground-truth data at multiple scales can effectively constrain initial permafrost conditions necessary for Earth System Model projections forced by Global Climate Model results for various emission scenarios, thus reducing the uncertainty in permafrost thaw predictions (McGuire et al., 2018).

Increasing air temperature translated to increasing ground temperature is the most noted and obvious driver of permafrost thaw. Changes in precipitation magnitude and seasonality can also influence ground temperatures (both positively and negatively), especially the timing and duration of snow cover that acts as an insulator to cold winter air temperatures (Jorgenson et al., 2010; Jafarov et al., 2018). Above-average summer precipitation can promote shallow thaw of warm permafrost as wet soils, with higher specific heat and bulk thermal conductivity, transfer thermal energy more effectively than dry soils. Douglas et al. (2020) report an average permafrost thaw rate of 0.7 $\pm$ 0.1 cm per cm increase in rainfall in boreal Alaska, with disturbed sites showing the greater thaw response to enhanced summer rain. Disturbance phenomena, such as fire (Brown et al., 2015; Minsley et al., 2016; Rey et al., 2020), flooding (Jepsen et al., 2016), vegetation change (Briggs et al., 2014), and infrastructure (Ghias et al., 2017) can further interact to modify the surface energy balance and shallow soil properties, which in turn influence permafrost dynamics, generally resulting in enhanced thaw rates in disturbed landscapes relative to their undisturbed counterparts. Wildfire is a prominent disturbance feature that shapes the carbon dynamics of northern ecosystems (Bond-Lamberty et al., 2007) by promoting permafrost thaw through the partial or complete combustion of the organic soil layer that provides insulation from warm summer temperatures as well as through the reduction in summer shading (Yoshikawa et al., 2002). Increases in wildfire activity and intensity have been documented and are projected to continue (Kasischke and Turetsky, 2006; Meredith et al., 2019), resulting in subsequent effects on permafrost thaw and lateral C export in both boreal and tundra regions (Gibson et al., 2018; Abbott et al., 2021; Ackley et al., 2021).

At the landscape scale, permafrost thaw responses to near-surface drivers typically display more spatial variability than that observed in the drivers themselves. This spatial mismatch in driver vs. response variability is due to prominent roles of landscape and subsurface characteristics, typically heterogeneous in nature, in mediating thaw rate and the coupled hydrologic and biogeochemical responses (Tank et al., 2020). **Table 1** describes positive, neutral, and negative associations between thaw likelihood/rate and various landscape and subsurface characteristics based on observational and modeling studies. Primary landscape features that influence thaw can generally be mapped *via* remote sensing methods, although ground-based measurements are needed for calibration and validation. In contrast, many influential subsurface characteristics cannot be determined remotely with current technology and thus require ground-based techniques such as coring or excavation to produce point-scale measurements. Geophysical techniques offer intermediate-scale characterization that can help bridge major gaps between point and remotely-sensed measurements (Minsley et al., 2012; Briggs et al., 2017). Tradeoffs between resolution and spatial coverage as well as tradeoffs between resolution and depth of investigation can be balanced through multi-method geophysical and remote sensing approaches (Walvoord and Kurylyk, 2016).

Active layer thickening, supra-permafrost talik formation, and complete permafrost loss generally show similar associations to thaw-influential factors shown in **Table 1** because these modes represent a continuum. Deeper subsurface properties become increasingly more relevant in the progression from active layer thickening to complete permafrost loss. Permafrost ice content has a negative association to most modes of thaw because greater volumetric ice content requires more thermal energy to complete the phase change to liquid water. Though the same thermodynamics apply to thermokarst processes, by definition, the development of thermokarst landforms relies on the existence of ice volume that exceeds the soil matrix pore space allowing for ground subsidence and/or slumping upon thaw. Accordingly, permafrost ice content has a positive association with thermokarst formation. Permafrost coverage is an important factor in determining propensity to the primary permafrost thaw modes. Consistent with the thaw state progression from active layer thickening to supra-permafrost talik development to complete loss beneath both terrestrial and aquatic systems, the least advanced thaw stage is more likely to be associated with areas with high (continuous) permafrost coverage whereas the more advanced stages are more common in areas with progressively less (discontinuous to sporadic) permafrost coverage.

HYDROLOGIC AND AQUATIC CARBON RESPONSES TO PERMAFROST THAW

Hydrologic susceptibilities and effects on aquatic C transport strongly depend on the mode and occurrence of thaw with respect to permafrost coverage (Striegl et al., 2007; Pokrovsky et al., 2015). Aquatic carbon biogeochemistry, transport and storage are tightly linked with water availability, flowpath and residence time. Permafrost C includes: soil organic carbon as particulate organic carbon (POC) and dissolved organic carbon (DOC); carbonate minerals; and dissolved inorganic carbon (DIC). Soil organic carbon receives the most attention because of its large magnitude and potential to eventually mineralize to the greenhouse gases CO_2 and CH_4 (Hugelius et al., 2014, 2020; Schuur et al., 2015). In the context of the water cycle, POC and DOC thawed from permafrost have three general fates depending on form, chemical composition and decomposability: (1) mineralization in-place or at some downgradient location; (2) lateral transport downgradient by surface and/or subsurface water; and (3) storage in-place or downgradient. Permafrost and seasonally frozen ground are barriers to vertical and horizontal subsurface water flow, so subsurface decomposition and transport diminish, and storage prevails when soil organic carbon is frozen. When surface runoff occurs, organic carbon export commonly derives from current vegetation sources with limited permafrost contributions (Dean et al., 2020; Beel et al., 2021). Dissolved inorganic carbon, the largest component of aquatic carbon exported to oceans, derives from mineralization of modern and permafrost organic carbon, exchange of CO_2 with the atmosphere, and dissolution (weathering) of carbonate minerals. DIC concentration is equal to the sum of the bicarbonate ion, carbonate ion, carbonic acid, plus dissolved CO_2 [$CO_{2(aq)}$] concentrations. For most streams, rivers, and lakes, bicarbonate is the prevalent species. Permafrost thaw makes frozen carbonates in permafrost available for

TABLE 1 | Influence of landscape and subsurface characteristics on permafrost thaw vulnerability to each primary mode of thaw.

Landscape and subsurface characteristics	Contribution to permafrost thaw − = negative; o = neutral; + = positive					Explanation	Support
	ALT	SPT	CPL	OT	TK		
Surficial landscape features							
Topographic gradient	+	+	+	o	+	Steeper subsurface hydraulic gradients enhance advective heat flow and promote thaw; thermokarst activity enhanced by erosion and mass transfer in areas of high relief.	Jorgenson et al., 2010; McKenzie and Voss, 2013; Evans and Ge, 2017
Organic layer thickness	–	–	–	o	–	Thicker organic layers provide insulation from warm summer air.	Park et al., 2013; Zhang et al., 2014; Walvoord et al., 2019
Surface water spatial coverage	+	+	+	+	+	Surface water provides a subsurface warming effect.	Wellman et al., 2013; Briggs et al., 2017; Rey et al., 2019
Snow cover thickness	+	+	+	o	+	Snow provides insulation to the subsurface from cold winter air.	Jafarov et al., 2018; Devoie et al., 2019; Walvoord et al., 2019
Wildfire occurrence	+	+	+	o	+	Higher intensity fires combust organics, thus reducing ground insulation from warm summer air.	Brown et al., 2015; Gibson et al., 2018; Rey et al., 2020
Thermally-independent subsurface factors							
Soil moisture content	+/–	+	+	o	+	Wetter soils provide less thermal insulation from warm summer air, which typically prevails over the latent heat evaporative cooling effect. However, high soil moisture in the AL for ice formation increases the required latent heat of fusion for thaw, slowing ALT.	Jorgenson et al., 2010; Devoie et al., 2019; Clayton et al., 2021
Shallow soil permeability (0–2 m depth)	+	+	+	o	–	Substrate with greater permeability allows more water to pass through and transfer heat, thereby enhancing thaw.	Wellman et al., 2013; Kurylyk et al., 2016
Deep soil permeability (>2 m depth)	o	+	+	+	o	Same as above, but applicable for deeper thaw processes.	McKenzie and Voss, 2013; Wellman et al., 2013
Thermally-dependent subsurface factors							
Permafrost spatial coverage	+	–	–	–	o	propensity of thaw mode with progressively decreasing permafrost coverage. The associations noted here relate to the	Synthesis of numerous studies
Depth to permafrost top	–	+	+	o	–	Greater depth to permafrost indicates more advanced state of thaw conditions.	Synthesis of numerous studies
Permafrost ice content	–	–	–	–	+	Greater ice content creates more thermal inertia, but is a key element in thermokarst processes.	Kokelj et al., 2013; Tank et al., 2020
Permafrost temperature	+	+	+	+	+	Warmer permafrost is more conducive to thaw than colder permafrost.	McKenzie and Voss, 2013; Rey et al., 2020
Permafrost total thickness	o	o	–	–	o	Thicker permafrost requires more time and energy to generate complete thaw.	Rowland et al., 2011

ALT, active layer thickening; SPT, supra-permafrost talik formation; CPL, complete permafrost loss; OT, open talik development; TK, thermokarst development. A negative (–)/neutral (o)/positive (+) association reflects the condition in which an increasing value for the factor in the left-most column generates reduced/unaffected/enhanced likelihood or rate of permafrost thaw via each primary mode.

downstream transport and/or weathering enhancing transport and downstream storage of particulate inorganic carbon (PIC) in stream and river channels and the production and export of DIC, primarily as bicarbonate. Mineralization to CO_2 is also a potential fate for thawed carbonates. These processes are observed at the river basin scale across the circumboreal region (Striegl et al., 2005, 2007; Tank et al., 2012a,b; Tank et al., 2016) and locally, especially near slopes having retrogressive thaw slumps (Zolkos et al., 2018). Anticipated trajectories of hydrologic and aquatic carbon responses to specified modes of permafrost thaw are described below and are summarized in **Table 2**.

Response to Active Layer Thickening

Thickening of the active layer increases potential shallow groundwater storage and promotes infiltration, thereby reducing runoff *via* overland flow and deepening subsurface flowpaths (**Table 2**, H1, H2). Collectively, these responses act to lengthen delivery times of water and solutes to stream and river networks (Ala-aho et al., 2018). This reduction in runoff mediated by

TABLE 2 | Summary of expected hydrologic and aquatic carbon responses to specified modes of permafrost thaw.

Response component	Expected response to modes of permafrost thaw − = reduced; o = neutral effect; + = enhanced					Explanation	Support
	ALT	**SPT**	**CPL**	**OT**	**TK**		
Hydrologic fluxes (H) contributing to aquatic systems							
H1: runoff as overland flow	–	–	–	**o**	**+/–**	Generally, permafrost thaw enhances the infiltration to runoff ratio; however, TK processes can enhance runoff by augmenting overall surface connectivity and effective surface channeling.	Connon et al., 2014; Koch et al., 2014; Bring et al., 2016
H2: active layer flow (summer)	**+**	**o**	**o/–**	**o**	**+/–**	Enhancement of active layer flow with ALT relies on water availability (recharge) and subsurface permeability.	Kurylyk et al., 2016; Walvoord and Kurylyk, 2016; Ala-aho et al., 2018; Evans et al., 2020
H3: seasonal variability (extremes)	–	–	–	**o**	**o**	Enhanced baseflow, with increasing relevance at larger scales from ALT to CPL, reduces seasonal variability in streamflow magnitude as well as temperature.	Smith et al., 2007; Karlsson et al., 2012
H4: perennial supra-permafrost flow	**o**	**+**	**o**	**o**	**o**	Perennial supra-permafrost flow is most affected by SPT and will depend on winter source/sink connectivity and lateral talik permeability. Aufeis formation provides one means of initial detection.	Jepsen et al., 2016; Walvoord et al., 2019; Terry et al., 2020
H5: sub-permafrost flow	**o**	**o**	**+**	**+**	**o**	CPL and OT enhance deep groundwater circulation and connectivity between shallow and deep, sub-permafrost water sources.	Rowland et al., 2011; Walvoord et al., 2012; McKenzie and Voss, 2013; Wellman et al., 2013; Evans et al., 2020
Carbon delivery (C) to surface waters							
C1: DOC age	**+**	**+**	**o**	**o**	**+**	Input from ancient permafrost C sources could increase apparent ^{14}C DOC age, especially in low-order streams in actively degrading watersheds.	Neff et al., 2006; Aiken et al., 2014; Wild et al., 2019; Schwab et al., 2020
C2: DOC export	**+/–**	**+/–**	–	**o**	**+/–**	Mobilization of permafrost DOC by ALT and SPT increases the potential for DOC export, particularly at local scales, but deeper flowpaths enhance the potential for OC degradation prior to stream delivery resulting in a mixed response especially across larger scales.	Striegl et al., 2005; Frey and McClelland, 2009; Abbott et al., 2015; Vonk et al., 2015b; Ackley et al., 2021
C3: flowpath length, depth, and residence time	**+**	**+**	**+**	**o**	**+/–**	In general, permafrost thaw opens and deepens subsurface flowpaths resulting in longer residences times; however, thermokarst processes may short-circuit pathways in some cases.	Frampton and Destouni, 2015; Ala-aho et al., 2018; Walvoord et al., 2019
C4: DIC export	**+**	**+**	**+**	**+**	**+**	Deeper and longer subsurface flowpaths increase the potential for OC mineralization and carbonate weathering enhancing DIC concentration and export depending on subsurface flow and biogeochemical conditions.	Striegl et al., 2005; Tank et al., 2012b, 2016, 2020; Vonk et al., 2015b; Zolkos et al., 2020b
C5: POC export	**o**	**o**	**o**	**o**	**+**	Mass wasting associated with some thermokarst processes enhances particulate delivery, including POC.	Abbott et al., 2015; Wild et al., 2019

ALT, active layer thickening; SPT, supra-permafrost talik formation; CPL, complete permafrost loss; OT, open talik development; TK, thermokarst development.

top-down permafrost thaw, including active layer thickening, generally promotes a reduction in seasonal streamflow variability as the subsurface acts as a low-pass filter for water delivery to streams (**Table 2**, H3). An exception to the pattern of reduced runoff to active layer thickening can occur in low-lying landscapes where thaw is accompanied by ground subsidence and surface channeling that increases surface hydrologic connectivity (Connon et al., 2014) and reduces transit time. Lateral supra-permafrost flow increases with active layer thickening unless the change is accompanied by a water table decline that causes water movement downward through a newly thawed zone with reduced permeability relative to material above (Koch et al., 2014; Walvoord and Kurylyk, 2016). Hydrologic susceptibility to active layer thickening is greatest in regions where subsurface flow is predominantly through the active layer, such as in continuous permafrost or in high-altitude discontinuous permafrost having shallow unfractured bedrock (Bring et al., 2016). Regions where much subsurface flow occurs *via* deeper flowpaths (i.e., taliks and sub-permafrost aquifers) are less impacted by subtle changes in active layer flow conditions.

Active layer thickening makes the uppermost centimeters of permafrost OC available for decomposition or transport as DOC. There is evidence that ancient DOC decomposes quickly following thaw, so a large fraction of DOC released by active layer thickening may mineralize near the source (Drake et al., 2015; Vonk et al., 2015a). In the few instances where radiocarbon ages have been measured on surface water DOC exported from permafrost terrain, aged DOC is detectable, particularly late in the thaw season (Neff et al., 2006; Striegl et al., 2007) but the signal is small and likely to diminish quickly due to high biolability (Vonk et al., 2013; Wickland et al., 2018) and/or dilution by modern C sources (Aiken et al., 2014; Dean et al., 2020; Schwab et al., 2020; **Table 2**, C1). Studies in northern permafrost regions indicate a mixed response in DOC riverine export to active layer thickening (Vonk et al., 2015b; **Table 2**, C2). Coupled hydrologic and biogeochemical conditions favoring increased export include release of slow-reacting recalcitrant DOC and rapid transit times through lateral subsurface flowpaths. However, increased downward transport of DOC from terrestrial and permafrost sources and longer subsurface residence times typically associated with active layer thickening (Frampton and Destouni, 2015; **Table 2**, C3) tends to enhance DOC consumption or sorption resulting in a net reduction of DOC export, particularly at larger scales (Striegl et al., 2005). Thawed conditions promote longer subsurface residence time for microbial mineralization of DOC, increased respiration, and increased DIC production through mineral weathering (**Table 2**, C4). This balance between DOC consumption and DIC production is confounded by vegetation changes associated with climate warming resulting in both increased DOC production by new vegetation and increased consumption of that DOC in the subsurface that alter the hydrologic DOC and DIC balance (Dornblaser and Striegl, 2015).

Response to Supra-Permafrost Talik Development

Hydrologic and biogeochemical susceptibilities for supra-permafrost talik development are similar to active layer thickening but include seasonal extension and moderation of flow variability (**Table 2**, H1–4 and C1–4), due to the potential year-round persistence of supra-permafrost flow even as winter ice forms above in the active layer. This allows for perennial reactivity and transport of C in the shallow subsurface through progressively deeper flowpaths (Walvoord et al., 2019). Increased seasonal flow duration can result in increased supra-permafrost delivery of DOC, DIC, CO_2(aq) and CH_4(aq) to inland waters, while relatively slow water movement in the subsurface also increases residence time for DOC mineralization and production of DIC and C-gases. The amounts of water and dissolved C that can be laterally transported during winter by supra-permafrost taliks depend strongly on talik permeability and the capacity of the system to maintain hydraulic gradients and subsurface connectivity to drive flow, conditions that are challenging to both characterize in the field and to represent in large-scale models. The presence of aufeis, or icings, can be useful in detecting substantial perennial supra-permafrost talik flow (Glass et al., 2020; Terry et al., 2020). The importance of supra-permafrost flow and the concomitant delivery or subsurface biogeochemical processing of permafrost C resulting from supra-permafrost taliks is a research area of required attention (Commane et al., 2017; Vonk et al., 2019). Plaza et al. (2019) directly measured carbon loss in permafrost soils over a 5-year warming experiment and pointed to lateral hydrologic export as a pathway potentially accounting for more than half of the large losses measured. Dissolved C transport through lateral taliks could help explain the large observed soil carbon losses. Likewise, Schwab et al. (2020) attribute a pulse of aged DOC, determined by radiocarbon activity in DOC, in the northern Mackenzie River basin of Canada, to lateral transport of thawed permafrost C through newly developed supra-permafrost taliks following several warm summer and winter seasons. The coupled hydrologic and biogeochemical role of lateral taliks remains poorly quantified and is likely to increase in importance with time and continued subsurface warming (Vonk et al., 2019; Walvoord et al., 2019).

Response to Complete Permafrost Loss

Areas of warm, thin, discontinuous permafrost are most likely to experience complete permafrost degradation. From an ecological perspective, complete loss of near-surface (∼3–4 m below the surface) permafrost is most pertinent, whereas from a hydrogeological perspective, complete permafrost loss (typically > 4 m) that establishes subsurface connectivity is most relevant. Complete permafrost loss has the potential to influence subsurface fluxes and flowpaths at intermediate to basin-wide scales by altering the hydrogeologic framework that controls groundwater flow, groundwater storage, and hydrologic connectivity among inland waters, shallow groundwater, and deep sub-permafrost groundwater. Widespread increases in streamflow minimums and river baseflow across the Arctic-boreal region suggest a multi-decadal shift from dominance of surface-water input toward increased groundwater input from a combination of connective supra-permafrost flow and sub-permafrost flow through open taliks (Smith et al., 2007; Walvoord and Striegl, 2007; Evans et al., 2020). This is consistent with the paradigm that proportionally more water is being recharged, stored, and circulated through deep subsurface pathways opened by extensive permafrost loss (Walvoord et al., 2012; Kurylyk and Walvoord, 2021). This conceptual model does not require enhanced precipitation or additional recharge from meltwater sourced by thawing permafrost or glaciers, but simply involves the rerouting of some surface water through deeper subsurface flowpaths, enhancing baseflow and reducing seasonal streamflow variability (**Table 2**, H1–5). Determination of the hydrologic impacts of complete permafrost loss requires site-specific consideration due to the strong influence of the underlying (unfrozen) hydrogeologic structure. For example, transitions from frozen to unfrozen coarse-grained sediment with high thawed permeability will invoke greater hydrogeologic changes than transitions from frozen to unfrozen bedrock with low thawed permeability. Anticipating broad changes in groundwater contributions to inland waters *via* permafrost loss and corresponding changes in flowpaths and water residence times will necessitate improved baseline hydrogeologic characterization of permafrost-impacted basins (i.e., airborne geophysics; Minsley et al., 2012) integrated with coupled permafrost and groundwater modeling. Some permafrost-impacted regions may demonstrate enhanced potential for groundwater resource development upon continued permafrost thaw (i.e., Lemieux et al., 2016).

Permafrost degradation over long timescales will affect aquatic C dynamics at all scales as hydrologic conditions shift from surface-water dominated to more groundwater influenced and residence times increase (Frey and McClelland, 2009; O'Donnell et al., 2012). Long-term consequences of permafrost thaw include increased downward movement of DOC from surface, shallow subsurface, and permafrost sources. This results in increased subsurface OC mineralization, rock weathering and bicarbonate production, enhancing DIC export to oceans (Striegl et al., 2005; Tank et al., 2012b; **Table 2**, C2, C4). Increased thaw and warmer soil conditions will likely also result in vegetation shifts and increased DOC production and export from non-permafrost sources (Tank et al., 2012a, 2016; Dornblaser and Striegl, 2015).

Response to Open Talik Development Beneath Water Bodies

Open taliks beneath large lake and river corridors within continuous and discontinuous permafrost regions can enhance surface water and sub-permafrost groundwater exchange (Wellman et al., 2013; **Table 2**, H5) and promote baseflow in large rivers throughout the winter. The direction of the exchange (drainage out or seepage in) depends on regional hydraulic gradients. For example, lakes that are topographically high on the landscape tend to be subjected to downward gradients and thus drain to the deeper groundwater system through open talik development, whereas topographically low lakes tend toward upward gradients that support seepage into the lakes through open talik development. However, local and regional gradients influenced by permafrost distribution are not well known without detailed field investigation, and hydraulic gradients can change in response to permafrost thaw. This complicates the expected hydrologic response associated with open talik processes.

Talik formation below water bodies exposes permafrost OC to decomposition and likely has consequences on lake CO_2 and CH_4 production and DOC composition. High proportions of anaerobic C mineralization near the thaw front of lake taliks have been observed (e.g., Heslop et al., 2020), with up to 80% of CH_4 lake emissions being associated with recent (decadal-scale) permafrost thaw (Walter Anthony et al., 2021). Lakes overlying thawed permafrost are known to have high CH_4 emission, largely by ebullition (Walter Anthony et al., 2021)

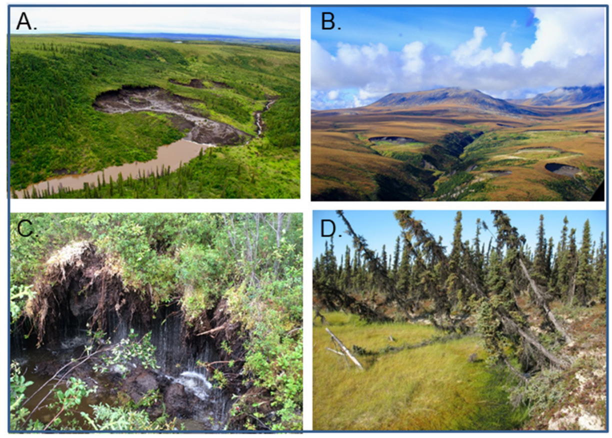

FIGURE 3 | Thermokarst processes affecting the storage and delivery of water and carbon across boreal landscapes. These include: **(A)** thaw slumps and enhanced particulate matter to stream input and **(B)** collapse and drainage features on the Peel Plateau, Northwest Territories, Canada, **(C)** active thermally-driven erosion pervasive in the Erickson Creek watershed, Alaska, USA, and **(D)** subsidence and bog formation in Innoko Flats, Alaska, USA. Photo credits: Scott Zolkos, Woodwell Climate Research Center **(A** and **B)**; Kim Wickland, U.S. Geological Survey **(D)**.

but other aqueous C consequences are not well characterized. Lake hydrological models suggest that increased subsurface connectivity among lakes will also lead to increased similarity in lake solute chemistry (Jepsen et al., 2016). Similar to other permafrost modes, open talik formation generally promotes deeper and longer subsurface flowpaths with increased time for OC mineralization and carbonate weathering, leading to enhanced DIC export (**Table 2**, C2–4).

Response to Thermokarst Formation

The numerous types of thermokarst development and complex geomorphic changes provide considerable challenge for the prediction of hydrologic responses. Thermokarst processes alter topography and erosional potential in moderate to steeply sloping watersheds and may lead to large mass wasting and dramatic increases in sediment and organic particulate transport to streams and rivers (**Figures 3A,B**). In relatively flat regions, localized thermal erosion (**Figure 3C**) and subsidence from thermokarst development may result in increases in surface water area (bog, pond, and lake development) (**Figure 3D**) (Walter Anthony et al., 2021) or loss of net surface water area (Nitze et al., 2018) due to increases in surface water drainage as thermokarst development evolves and the surface water connectivity increases (Connon et al., 2014). Episodic thermokarst events can result in the formation of new lakes or loss of old lakes in days to weeks, dramatically altering the ecosystem services associated with those waterbodies (Nitze et al., 2020).

Thermokarst development can have a dramatic and visual effect on aquatic C dynamics. Mass wasting of hillslopes and shorelines contributes large amounts of POC, aged DOC, nutrients and sediment with short residence times into lakes and streams (**Figure 3A**; Abbott et al., 2015; Wild et al., 2019; **Table 2** C1–C3, C5). These inputs tend to be biogeochemically reactive on short time scales affecting aquatic ecosystem respiration, primary production, mineral weathering, greenhouse gas emissions and other processes (Vonk et al., 2015a,b; Walter Anthony et al., 2021). Subsurface organic C buried and frozen for centuries to millennia can be respired quicky to enhance stream CO_2 emissions when exposed by retrogressive thaw slumps (Zolkos et al., 2019). Countering this process, carbonic acid weathering of carbonate minerals exposed by thermokarst and erosion decreases stream and lake CO_2 emissions while increasing bicarbonate production and transport (Tank et al., 2016; Zolkos et al., 2018, 2020b; **Table 2**, C4). This pattern is complicated in regions where sulfuric acid weathering of carbonates produces CO_2 (Zolkos et al., 2018), emphasizing the need to understand local geological and geochemical conditions when assessing the role of permafrost thaw on aquatic carbon cycling. Carbon storage is a component of POC delivery to inland waters *via* thermokarst processes. POC released by thermokarst can be transported downstream to coastal regions, lakes, or wetlands where duration of storage is a function of POC biodegradability and oxygen conditions in bottom sediments at the relocation sites (Beel et al., 2020).

Accounting for Competing Influences and Heterogeneity in Permafrost C Response to Thaw

Because of the importance of changing hydrologic conditions in thawing permafrost landscapes as described in this section, the primary pathways of permafrost carbon and trajectories of C export (**Table 2**) depend on joint consideration of factors influencing subsurface water flux (permeability, gradient, hydrologic connectivity, and sustained water source) (**Figure 4A**) and carbon source (chemical composition, amount, and OC degradability) (**Figure 4B**), in addition to factors that control modes and rates of permafrost thaw, described in section Factors Influencing Vulnerability to Permafrost Thaw (**Table 1**, **Figures 4C,D**). Challenges in assessing permafrost C pathways (**Figure 4E**) arise because landscape and subsurface

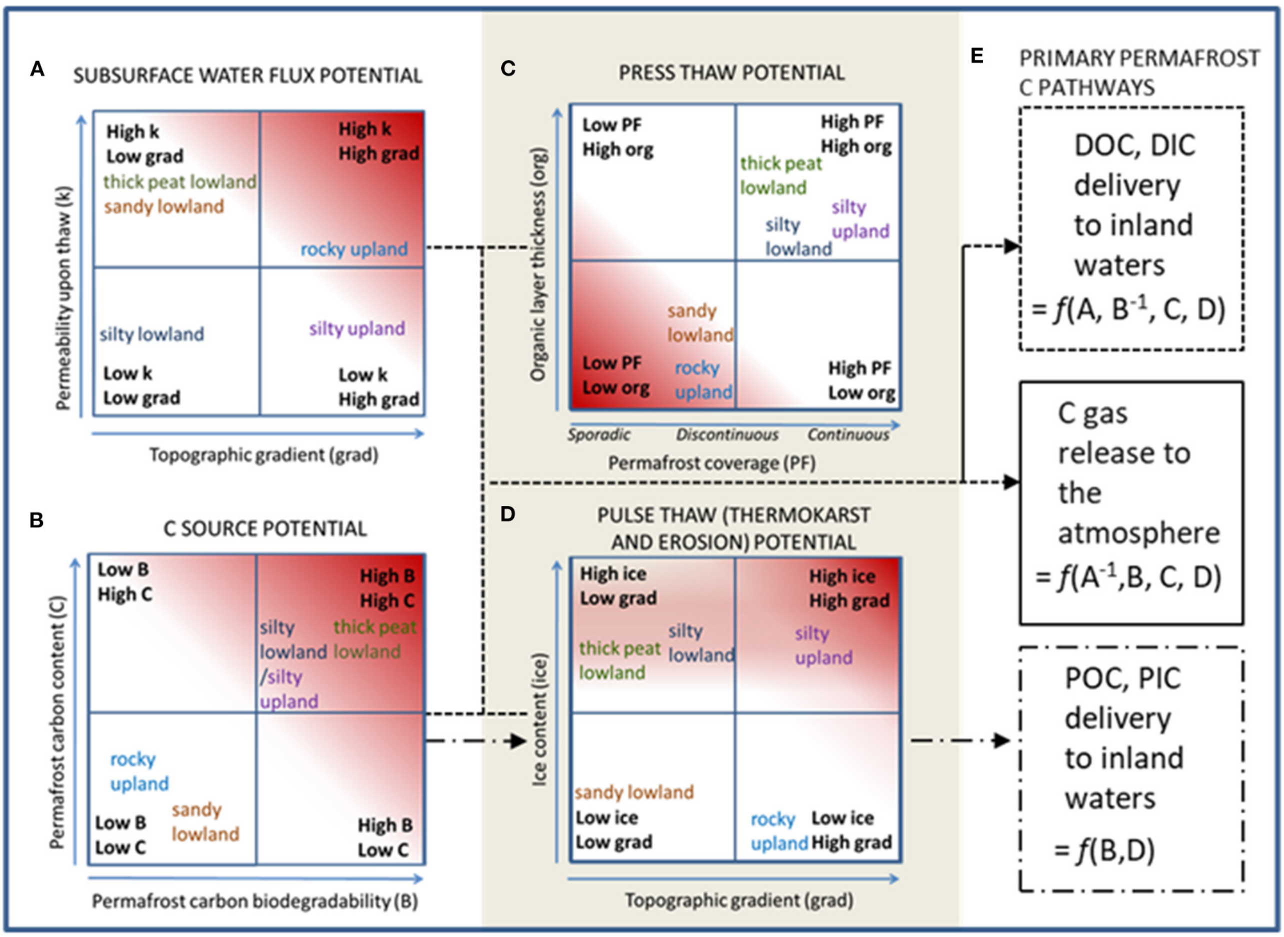

FIGURE 4 | Framework depicting primary characteristics influencing **(A)** subsurface hydrology **(B)** permafrost carbon source strength, and **(C,D)** permafrost thaw, and the collective effect on **(E)** the fate of permafrost carbon delivered to coastal and inland waters where it is mineralized and/or stored in sediments. Red shading represents conditions with the highest potential for **(A)** lateral hydrologic flux, **(B)** carbon source strength, **(C)** press thaw, and **(D)** pulse thaw. Representative, generalized hydrogeologic landscapes (thick peat lowland, sandy lowland, silty upland, silty lowland, rocky upland) are noted in colored text and are provided to demonstrate the non-alignment of hydrologic, biogeochemical, and thaw potentials.

characteristics that influence subsurface water flux, carbon source, pulse and press thaw (**Figures 4A–D**) are not commonly aligned. For example, as illustrated in **Figure 4**, systems that have among the greatest potential carbon source characteristics, such as low-lying thick peatlands and silty uplands and lowlands (Vonk et al., 2015a,b; Wickland et al., 2018), have limited potential for lateral subsurface water flux upon thaw due to low hydraulic gradient or low thawed permeability (Kurylyk et al., 2016; Ebel et al., 2019). Similarly, rocky or sandy uplands, with the greatest potential for lateral subsurface hydrologic flux upon thaw (Evans and Ge, 2017; Ebel et al., 2019), have limited carbon source potential (Saito et al., 2020). Incorporation of joint potentials is thus required for estimating the magnitude of the pathways for permafrost carbon fate *via* lateral dissolved or particulate pathways or release to the atmosphere as gas (**Figure 4E**, **Table 2**). Evidence for enhanced subsurface connectivity and lateral flowpath development accompanying thaw in discontinuous permafrost landscapes continues to mount (Connon et al., 2018; Parazoo et al., 2018; Rey et al., 2019) suggesting an increasing potential for lateral hydrologic C transport. Models that do not consider lateral hydrologic C flux may underestimate dissolved C delivery to inland waters in response to permafrost thaw and overestimate C gas released to the atmosphere.

Additional challenges toward the goal of projecting hydrologic and biogeochemical change in northern latitudes arise from the substantial degree of spatial heterogeneity in landscape and subsurface characteristics that give rise to the discontinuous permafrost distribution typical of these regions. Consequently, aquatic C and hydrologic responses to climate change in discontinuous permafrost also tend to display a high degree of heterogeneity (Tank et al., 2020). Furthermore, these regions are subject to all five primary modes of thaw, thus elevating the complexity of conceptual model representation. The nature of discontinuous permafrost and thaw evolution in these systems heightens the impact of subsurface flow and C transport relative to continuous permafrost where water movement and carbon cycling are mainly confined to the active layer. The impact of continued thaw toward sporadic and isolated permafrost coverage lessens as the capacity to affect hydrologic and biogeochemical processes and pathways becomes increasingly diminished. Thaw occurring in regions near the continuous-discontinuous permafrost transition is increasingly recognized as having the highest potential for hydrogeologic impact as subsurface connectivity evolves non-uniformly (Rey et al., 2019). Computational limitations and current characterization gaps have justified major simplification of subsurface lateral flow processes in large-scale Earth System Models (ESMs) used to predict changes in water and carbon dynamics of thawing permafrost regions. Major advances must overcome these challenges. Establishing guidelines for the required spatial and vertical resolution of influential landscape and subsurface characteristics controlling thaw and its hydrologic and biogeochemical impact in diverse Arctic and boreal regions is an important step in this direction.

CONCLUSIONS

Carbon cycling is largely driven by water availability, phase and movement. As permafrost thaws, water and carbon become available for increased biogeochemical processing and transport *via* deeper subsurface pathways. Locations, landscape and subsurface factors, and modes of thaw will determine the ultimate consequences of thaw with complexities arising in part because characteristics influencing thaw are not necessarily in line with characteristics influencing coupled hydrologic and biogeochemical responses. Shallow permafrost thaw processes (active layer thickening and thermokarst formation) have proportionally large hydrologic and C-cycle impacts in systems with continuous permafrost coverage, whereas systems having discontinuous to sporadic permafrost coverage tend to be progressively more impacted by deeper permafrost thaw processes, including supra-permafrost talik development and complete permafrost loss. A major challenge for projecting the fate and transport of northern permafrost C in the twenty-first century and beyond is addressing change resulting from permafrost thaw in water flowpaths, fluxes, and availability. Vulnerability depends not only on carbon source (composition, abundance, reactivity) and likelihood of permafrost thaw, but also on factors and drivers sensitive to subsurface water movement, hydrologic connectivity, and the availability of water to drive and sustain flow and the transport of dissolved and particulate C. Improving projections of permafrost C release for future planning in northern boreal regions requires strong, general understanding of the timescales associated with various modes of permafrost thaw, a framework describing observable factors in the context of permafrost thaw likelihood (**Table 1**), joint hydrologic and biogeochemical considerations (**Figure 4**), and understanding of the expected trajectories of hydrologic and aquatic C responses to thaw (**Table 2**). Yet, critical research questions remain in determining spatial and vertical resolution thresholds for landscape and subsurface characterization (**Table 1**) required to adequately represent critical thermal, hydrologic, and biogeochemical processes in models designed to assess and predict permafrost thaw and its impacts. Rapid advances in remote sensing are being made to improve relevant near-surface characterization. Deeper subsurface characterization over large scales is a critical research need and current data gap. It follows that predictive large-scale permafrost models that focus on near-surface permafrost most directly apply to modes of thaw occurring in the shallow subsurface (active layer thickening and thermokarst formation). Incorporation of deeper thaw processes in large-scale predictive models to represent supra-permafrost and open talik development are needed for comprehensive evaluation of coupled surface water and groundwater resources in transitional permafrost environments (McKenzie et al., 2021). Comprehensive evaluation of the fate of circumboreal permafrost carbon and associated climate feedbacks in Earth System Models requires the incorporation of permafrost-hydrologic interactions that influence terrestrial and aquatic C cycles and lateral C transport to inland waters and coastal oceans in addition to land and water surface exchange of C with the atmosphere.

Permafrost thaw and its impacts to the water and carbon cycles are of interest not only to earth scientists, local communities, and land managers, but also to economists, military advisors, and policy makers due to the important role of northern high latitudes in the future global economy and international relations. Improved understanding and prediction of permafrost thaw impacts to hydrology will help mitigate economic losses derived from climate change (Melvin et al., 2017; Yumashev et al., 2019). Cost-effective development of infrastructure and groundwater resources in thawing permafrost regions also rely on a solid understanding of plausible futures.

This synthesis broadly applies to northern permafrost regions and provides a framework for discussing the implications of climate change and permafrost thaw on hydrologic and aquatic C dynamics across diverse landscapes. Understanding the underlying complexities that shape the hydrologic and aquatic C responses to various modes of permafrost thaw will elevate confidence in their prediction from the low to medium levels noted in the IPCC 2019 report.

AUTHOR CONTRIBUTIONS

MW led the manuscript writing with significant contributions from RS. Both authors contributed to the article and approved the submitted version.

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The Mittimatalik Siku Asijjipallianinga (Sea Ice Climate Atlas): How Inuit Knowledge, Earth Observations, and Sea Ice Charts Can Fill IPCC Climate Knowledge Gaps

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The IPCC *special report on the ocean and cryosphere in a changing climate* (SROCC) highlights with high confidence that declining Arctic sea ice extents and increased ship-based transportation are impacting the livelihoods of Arctic Indigenous peoples. Current IPCC assessments cannot address the local scale impacts and adaptive needs of Arctic Indigenous communities based on the global, top-down model approaches used. Inuit maintain the longest unrecorded climate history of sea ice in Canada, and to support Inuit community needs, a decolonized, Inuit knowledge-based approach was co-developed in the community of Mittimatalik, Nunavut (Canada) to create the Mittimatalik siku asijjipallianinga (sea ice climate atlas) 1997–2019. This paper presents the novel approach used to develop the atlas based on Inuit knowledge, earth observations and Canadian Ice Service (CIS) sea ice charts, and demonstrates its application. The atlas provides an adaptation tool that Mittimatalik can use to share locations of known and changing sea ice conditions to plan for safe sea ice travel. These maps can also be used to support the safety and situational awareness of territorial and national search and rescue partners, often coming from outside the region and having limited knowledge of local sea ice conditions. The atlas demonstrates the scientific merit of Inuit knowledge in environmental assessments for negotiating a proposal to extend the shipping seasons for the nearby Mary River Mine. The timing and rates of sea ice freeze-up (October–December) in Mittimatalik are highly variable. There were no significant trends to indicate that sea ice is freezing up later to support increased shipping opportunities into the fall. The atlas shows that the first 2 weeks of November are critical for landfast ice formation, and icebreaking at this time would compromise the integrity of the sea ice for safe travel, wildlife migration and reproduction into the winter months. There was evidence that sea ice break-up (May–July) and the fracturing of the nearby floe edge have been occurring earlier in the last 10 years (2010–2019). Shipping earlier into the break-up season could accelerate the break-up of an already declining sea ice travel season, that Inuit are struggling to maintain.

Keywords: Indigenous knowledge, Inuit Qaujimajatuqangit, decolonizing research, research co-production, sea ice, climate change adaptation, Arctic

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INTRODUCTION

The International Panel on Climate Change (IPCC) Special Report on the Ocean and Cryosphere in a Changing Climate (SROCC) outlines that between 1979 and 2018, sea ice in the Arctic decreased by ∼13% per decade (IPCC, 2019, p. 6). This decline in sea ice is expected to continue into the mid-century having significant impacts on Arctic Indigenous peoples nutritional, cultural, and overall health and well-ness (IPCC, 2019, p. 15). Inuit communities are already dealing with dangerous sea ice travel conditions, limiting access to critical hunting locations and country food sources, and causing high rates of search and rescue, injury, trauma, and tragic deaths (Durkalec et al., 2014; Clark et al., 2016a,b; Driscoll et al., 2016; Kenny et al., 2018b; Ford et al., 2019). Additionally, the surge in shipping activity as a result of changing ice conditions is also impacting Arctic Indigenous peoples (IPCC, 2019). In the Canadian Arctic there has been a three-fold increase in the distance traveled by ships between 1990 and 2015 (Pizzolato et al., 2014, 2016; Dawson et al., 2018). This exposes Indigenous coastal communities to a higher risk of accidents, pollution, noise, invasive species, and disruptions to subsistence hunting areas, wildlife reproduction, populations, and migration routes (Huntington et al., 2015; ICC-Alaska, 2015; Meredith et al., 2019).

IPCC assessments are limited in addressing the climate change questions of Arctic Indigenous communities because of the global scale used in predictive models. Also, the top-down, model-focused approaches used by a majority of assessments are also a barrier to addressing the specific sea ice climate change adaptive needs of Arctic communities (Ford et al., 2012). Inadequate supports to engage meaningfully with Indigenous peoples limits an understanding of the cumulative impacts of colonialism and climate change on Arctic Indigenous communities (Ford et al., 2012; Cameron et al., 2015; IPCC, 2019, p. 15). For example, increased shipping and changes to on-ice travel are not mutually exclusive impacts. In the Inuit community of Mittimatalik (Nunavut, Canada; **Figure 1**), shipping and on-ice travel are in direct conflict with one another.

Sikumiut are a committee of Inuit sea ice users that govern the SmartICE community-based sea ice monitoring program (http://www.smartice.org/ice-safety) in Mittimatalik. Sikumiut members wanted to be able to share with younger generations where and when the sea ice is changing to support safer on-ice travel. Sikumiut also wanted to investigate the potential impacts of a proposed extension to the shipping season by Baffinland Iron Mines (BIM), the company that operates the Mary River iron ore mine and port near the community (**Figure 1**). Sikumiut are concerned about BIMs proposal to ship earlier during sea ice break-up and later as the sea ice is freezing. The nearby sinaa (floe edge), a stable landfast sea ice edge critical for community hunting, is highly anticipated during the freeze-up season. Avoiding disturbances to the sinaa and tuvaq (landfast ice) as they form is critical to community members for safe sea ice travel throughout the season, as well as for wildlife habitat and migration.

This collaborative research project with Sikumiut began in 2017. In earlier phases of our work sea ice travel safety maps for the winter and spring travel seasons were developed based on Sikumiut's Inuit Qaujimajatuqangit (Wilson et al., in press). Inuit Qaujimajatuqangit (IQ) is commonly used to describe Inuit knowledge, but it also encompasses all aspects of Inuit "values, world-view, language, social organization, knowledge, life skills, perceptions, and expectations" (Government of Nunavut and Nunavut Department of Education, 2007). As a result, these IQ-based sea ice maps share more than locations of safe and hazardous ice conditions. Embedded in the Inuktitut place names and sea ice terms are important information for sea ice travel and survival (Wilson et al., in press). These Sikumiut sea ice IQ travel safety maps also provide a time-integrated baseline of the winter and spring sea ice travel conditions for Mittimatalik. Typically, meteorologists call these baselines "climatologies," comprising databases of historical weather or sea ice observations (WMO, 2017). These climatologies are used to compare and track changes over time, and are used particularly to monitor climate change trends. Sikumiut's IQ-based sea ice climatology is maintained by passing down their IQ through generations, and orally sharing their extensive and recent travel experiences on the sea ice. Sikumiut's sea ice climatology is therefore not in a database, but exits in the collective minds of these expert sea ice travelers. Also, their climatology is not focused on ice conditions in a general scientific sense, but more specifically on ice conditions supporting safe travel and spatio-temporal patterns of ice features that support hunting. To support Sikumiut's climate change adaptation needs, a novel approach was co-developed to document for the first time their sea ice IQ to create the Mittimatalik siku asijjipallianinga (sea ice climate atlas).

The goals of this paper are three-fold. First, we outline the unique IQ-based research co-production approach that utilized earth observations and Canadian Ice Service (CIS) sea ice charts to create a sea ice climatology for the community of Mittimatalik. We present how Sikumiut's IQ was the foundation for the development, analysis and production of the final maps in the siku asijjipallianinga. Second, we present the utility of the atlas in summarizing Mittimatalik's sea ice trends (averages, variability, spatial changes) over the 23-year climatological period (1997–2019). Third, we demonstrate the value of such IQ-based, community-scale sea ice climatologies for local and regional scales.

This paper does not include an analysis of the atmospheric drivers for local sea ice change in Mittimatalik. This would normally accompany the presentation of a regional sea ice climatology, but this was not requested by Sikumiut. Also, this paper is not an example of integrating or incorporating IQ into western science. In this IQ-based sea ice climatology, we utilized other data sources to address Inuit specific research questions. This paper provides an example of an IQ-based research co-production approach in practice, including supplementary data sources, to fill the climate knowledge gaps and support adaptation needs for the community of Mittimatalik.

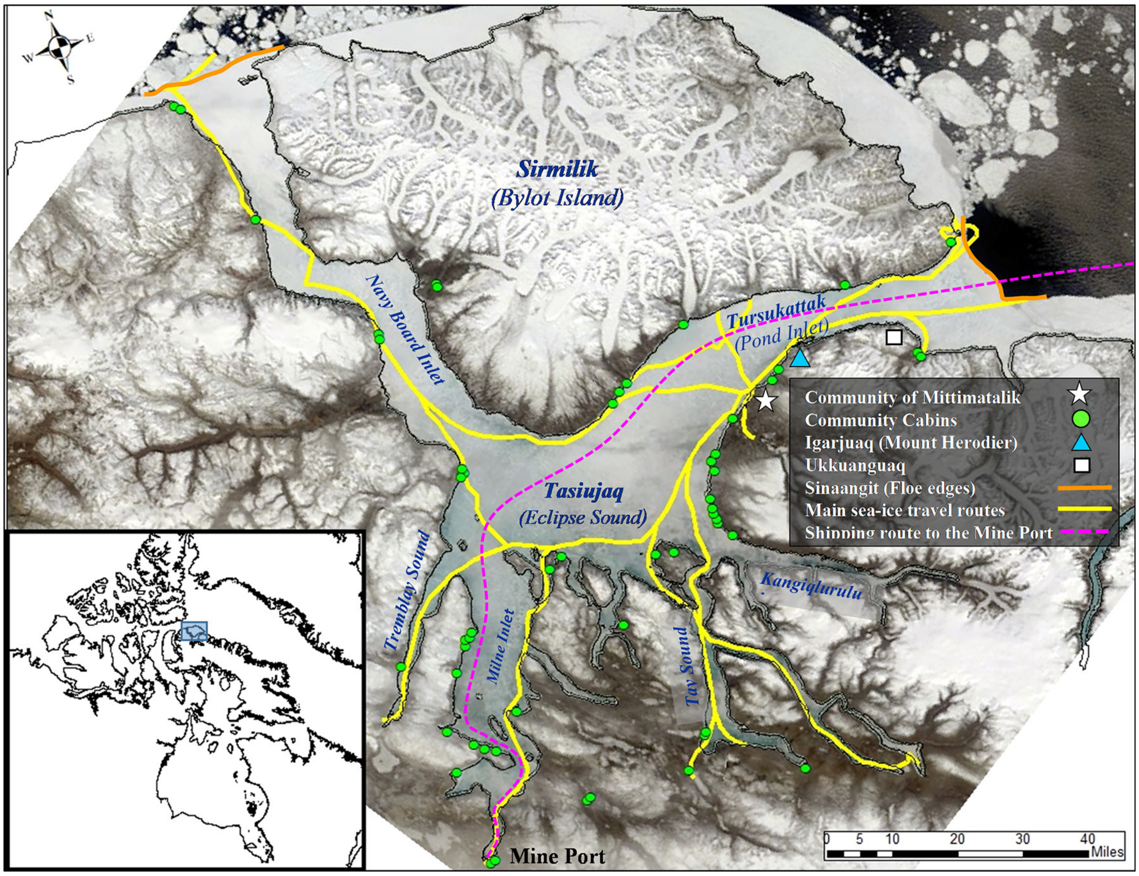

FIGURE 1 | Geographical location of the community of Mittimatalik, Nunavut, Canada. Background image MODIS True Color Composite, June 9, 2019 (NASA, 2019).

BACKGROUND

In this background section we briefly review the impacts of climate change and colonialism on safe sea ice travel across Inuit Nunangat. Inuit Nunangat is the Inuit homeland in Canada that covers the four Inuit land claim settlement regions of: Inuvialuit Settlement Region (Northwest Territories), Nunavut, Nunavik (northern Québec), and Nunatsiavut (northern Labrador) (ITK, 2018). We also present the Inuit community of Mittimatalik, outline our 6-year research co-production journey, introduce the research partners and co-authors, and how the need for a Mittimatalik sea ice climatology evolved. Finally, we review the current information sources available to build sea ice climatologies at community scales in the Canadian Arctic. In this paper we use the Mittimatalik Inuktitut sea ice and geographic terms and **Table 1** has been provided for reference to the equivalent English terms while reading.

Climate and Colonial Impacts for Safe Sea Ice Travel

The IPCC SROCC defines climate as the "average weather … over a period of time ranging from months to thousands or millions of years" (Portner et al., 2019, p. 680). In Inuktitut there is no word for climate or climate change. The closest word to climate in Inuktitut is *sila,* which has been defined as weather and the spiritual power that controls weather (Fox, 2004; Leduc, 2007). In Inuktitut, the term *silaup qanuinnirigajuktanga* is now used for climate and the direct translation from Inuktitut is "[t]he usual temperature, rain or snow and wind conditions of an area over a very long number of seasons" (GN and NTI, 2005, p. 39). Climate change is defined as "A change in the state of the climate that can be identified (e.g., by using statistical tests) by changes in the mean and/or the variability of its properties and that persists for an extended period, typically decades or longer" (Portner et al., 2019, p. 680). The Inuktitut term *silaup asijjiqpallianinga* is the term used for climate change and has various definitions that include: "A difference in the usual and extreme global temperatures that is not just a short cycle, but lasts for decades" (GN and NTI, 2005, p. 35); and the "ongoing and continuous change in sila" (Cameron et al., 2015, p. 278). The Inuktitut term sila is much more nuanced. For the context of this paper we are using the Government of Nunavut and Nunavut Tunnagavik Inc., Definitions (2005), but for a more in-depth discussion see (Fox, 2004; Leduc, 2007; Cameron et al., 2015).

TABLE 1 | Mittimatalik Inuktitut sea ice terms and geographic place names with English equivalent terms and definitions.

Inuktitut term	English equivalent
Aajuraq	Lead (singular). A crack in the sea ice that gets wider in the spring and is not always possible to cross
Aajurait	Leads (plural). Cracks in the sea ice that gets wider in the spring and are not always possible to cross
Imaqainnaujattuq ukiutamaa	Water that runs from glaciers onto the sea ice and melts it.
Ivujuk	Ridges, high areas of rough ice you have to travel around
Mittimatalik	Pond Inlet
Mittimatalingmiut	People of Mittimatalik
Nagguti	A crack in the ice that refreezes in winter. Narrow enough to cross but can be dangerous
Naggutiit	Cracks in the ice that refreeze in winter. Narrow enough to cross but can be dangerous
Sila	Weather and climate
Silaup qanuinnirigajuktanga	Climate
Silaup asijjiqpallianinga	Climate change
Siku	Sea ice
Siku asijjipallianinga	Changes to the sea ice (sea ice atlas)
Sikumiut	People of the sea ice, self-titled name of the Inuit management committee that governs the SmartICE community-based sea ice monitoring program (smartice.org) in Mittimatalik
Siku saattuq aragulimaamik	Thin ice all year
Siku saattuq upingaat pigiarningani	Thin ice in spring
Sinaa	Floe edge (singular)
Sinaangit	Floe edges (plural)
Sirmilik	Bylot Island. The place of glaciers
Tasiujaq	Eclipse Sound marine region
Tursukattak	Pond Inlet marine region
Tuvaq	Landfast sea ice

Environmental changes to sea ice travel is having profound impacts on the physical, cultural, and mental health of Inuit (Cunsolo Willox et al., 2013; Ford et al., 2013b; Durkalec et al., 2015; Pearce et al., 2015). Sea ice provides a stable platform to access country food (wild food from plants and animals, which is gathered and caught from the land and ocean). Changing weather and sea ice conditions are limiting Inuit access to critical hunting locations and country food sources (Laidler et al., 2009; Clark et al., 2016a; Kenny et al., 2018b). The high cost of store-bought foods in Inuit Nunangat means Inuit food insecurity rates are eight times higher than the rest of Canada (Kenny et al., 2018a). Inuit are now having to navigate new, longer, and more dangerous routes on the sea ice to access country food, which increases the risk of becoming lost in unfamiliar areas. Changes to traditional sea ice routes have also led to the use of more fuel, running out of gas, breaking through unexpected areas of thin ice, having to travel over rough ice and/or land resulting in snowmobiles and other equipment being lost and damaged (Ford et al., 2007; Durkalec et al., 2015; Driscoll et al., 2016; Fawcett et al., 2018). Search and rescue requests have not only increased due to changing weather and sea ice conditions, but also due to mechanical breakdown and running out of gas (Durkalec et al., 2014; Clark et al., 2016a). Rates of unintentional injury and trauma are extremely high in Inuit Nunangat, and in Nunavut specifically they "are more than twice the national average… and the leading cause of morbidity and mortality" (Durkalec et al., 2014; Clark et al., 2016a, p. 44).

As identified in the IPCC SROCC, climate change has left some experienced hunters doubting their weather and sea ice forecasting skills (IPCC, 2019); however, many hunters still have confidence in their IQ to navigate and make critical decisions on the sea ice, even under changing sea ice conditions (Gearheard et al., 2006; Pearce et al., 2010; Wilson et al., in press). The high rates of sea ice related injury and search and rescue experienced by Inuit are not simply due to climate change, but are intertwined with the ongoing effects of colonialism that have weakened the transmission of sea ice IQ through reduced language and practice (Tester and Kulchyski, 1997; Damas, 2002; MacDonald, 2018). The transition of Inuit into settlements, wage labor, and residential schools resulted in generations of Inuit being deprived of the time and access to the sea ice to develop this IQ through observations and experiences with parents and Elders (Tester and Kulchyski, 1997; Damas, 2002; QIA, 2014; TRC, 2015; MacDonald, 2018). Colonialism has left some Inuit unable to communicate in Inuktitut, impacting their ability to learn, understand, and share sea ice conditions and experiences with hunters and Elders (Heyes, 2011; Pearce et al., 2011; Ford et al., 2013a). Despite these challenges, sea ice IQ has endured and continues to be gained through experience and practice. Inuit continue to share their sea ice observations and knowledge to make safe sea ice travel decisions (Pearce et al., 2010; Ford et al., 2013a; Gearheard et al., 2013; ICC-Canada, 2014; Wilson et al., in press).

Evolution of the Research Partnership and Project

The community of *Mittimatalik* (Pond Inlet) is located at the northern tip of Baffin Island in Nunavut (**Figure 1**). It has a population of ∼1,600 people, of which 92% are Inuit and speak Inuktitut as their first language (Statistics Canada, 2017). The sea ice around the community begins to freeze in late October, and is normally safe for travel by mid-November once the ice becomes tuvaq (land-fast ice or stable sea ice that is frozen to the land) (Wilson et al., in press). *Mittimatalingmiut* (people of Mittimatalik) travel on the sea ice to hunt and fish for country food (caribou, narwhal, beluga, seal, and charr) and to spend time away from town at family cabins. Areas commonly traveled around Mittimatalik discussed in this paper include: Navy Board Inlet, *Tasiujaq* (Eclipse Sound), and *Tursukattak* (Pond Inlet; **Figure 1**). There are two sinaangit (plural of sinaa = floe edges) in the region, one at the entrance to Navy Board Inlet and one at the entrance to Tursukattak (**Figure 1**). Sinaangit are stable edges of tuvaq, located beside areas of open water that remain clear of ice throughout most of the sea ice season. The Tursukattak sinaa is located ∼65 km from the community and is one of the main hunting and fishing locations that Mittimatalingmiut use from December to early July.

Mittimatalingmiut want to maintain their sea ice travel, and are looking to additional information sources to augment their decision-making. Some members of the community heard about SmartICE and invited Trevor Bell to Mittimatalik in 2015 to discuss how SmartICE could support the community's sea ice travel safety concerns. SmartICE (smartice.org) is a work integration social enterprise that provides ice thickness measurements for Inuit communities using: *in-situ* instruments (SmartBUOYs) located at strategic travel locations on the sea ice; and a mobile sensor (SmartQAMUTIK) towed behind a snowmobile throughout the season on the main sea ice trails (Bell et al., 2014). Bell and Katherine Wilson spent 2 years developing relationships and trust to establish an Inuit-led SmartICE operations team in Mittimatalik. Bell is a co-author on this paper, a co-supervisor for Wilson, and the founder of SmartICE. Wilson, the lead author of this paper, is a PhD candidate with Memorial University of Newfoundland. She is also an employee of the Government of Canada for over 25 years, currently with the Canadian Ice Service (17 years in total), part of Environment and Climate Change Canada (ECCC). Wilson returned to school in 2015 under the co-supervision of Bell and Gita Ljubicic (McMaster University, also co-author), to retrain in decolonizing and Indigenous research approaches, and to put into practice a different way of doing research that empowers Inuit self-determination (Wilson et al., 2020).

Andrew Arreak, co-author, lives in Mittimatalik and was hired and trained in 2015 as the SmartICE community operator, now the Nunavut Operations Lead for the Qikiqtaaluk North (Baffin) region of Nunavut. In 2016, a 10-person management committee of Elders, experienced sea ice users and youth was established to govern SmartICE in Mittimatalik. Sikumiut, which means "people of the sea ice" in Inuktitut, is the self-titled name of the management committee (also co-authors on this paper, see Acknowledgments for list of members). Over these initial 2 years, Sikumiut began to share their concerns with Bell and Wilson about previous research relationships and younger Inuit lacking the necessary IQ needed to travel safely on the sea ice.

In 2017, our third year working together, we spent time planning the research focus and co-developing a cross-cultural research approach, called the Sikumiut Model (Wilson et al., 2020), with six goals:

1. Support Inuit self-determination in research;
2. Embrace Inuit decision-making;
3. Prioritize community-based research needs;
4. Develop Inuit specific values for research;
5. Strengthen Inuit youth capacity; and
6. Change the role of non-indigenous research partners.

In the Sikumiut Model, the research is focused on community research needs and building Inuit youth capacity in research. As a result, we worked to change the status quo, and the role of the non-Indigenous research partners was reconceptualized as facilitators and mentors to train Inuit youth in Mittimatalik to do this research themselves. Arreak was hired as the Inuit youth researcher to work on this project alongside his part-time SmartICE duties. To formalize the co-produced research approach, an agreement between Sikumiut and Memorial University was developed, which outlined the project goals, as well as roles and responsibilities of the Inuit and non-Indigenous project partners (Wilson, 2018). The research agreement also specified that the knowledge and data from this project are owned by Sikumiut, and they gave consent to Wilson to publish the results as part of her PhD requirements.

FIGURE 2 | (A) Sikumiut members mapping their sea ice IQ, November 2018. Photo credit Katherine Wilson. **(B)** Sikumiut members reviewing the Mittimatalik siku asijjipallianinga maps, March 2021. Photo credit Shelly Elverum.

In 2018 we began the research phase of the project. Sikumiut wanted to first document and share their sea ice IQ with the next generation to improve safe sea ice travel in the community. During 2018, workshops were held to document Sikumiut's sea ice terminology and to map Sikumiut's knowledge of safe and dangerous sea ice travel areas from winter to early summer as the sea ice is breaking up (**Figures 2A**, **3**). Between 2019 and 2021

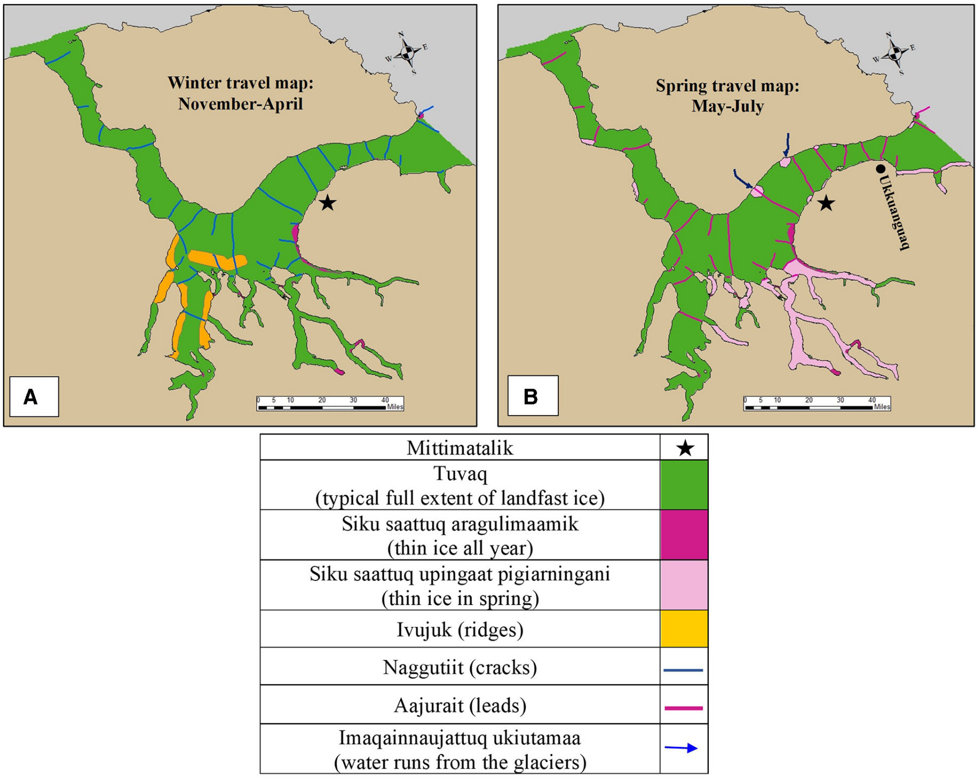

FIGURE 3 | Sikumiut seasonal sea ice safety travel maps **(A)** Winter sea ice IQ travel map, November to April; **(B)** spring sea ice IQ travel map, May to July. See on-line versions at http://www.smartice.org/ice-safety.

this sea ice IQ was made accessible to the community through the development of a sea ice terminology booklet, posters and seasonal maps of safe and dangerous areas to travel (Wilson et al., in press).

Over several meetings Sikumiut members discussed that while the sea ice freezes and breaks up differently each year, changes in sea ice conditions are now beyond what they would consider normal. Sikumiut members were interested in understanding where the sea ice was becoming more dangerous, so they could adapt their travel routes to maintain their hunting and fishing activities. In addition, Sikumiut were also concerned about a request from BIM to extend the shipping to/from the mine into the sea ice season. **Figure 1** shows the current shipping route from Baffin Bay, past the community, into Tasiujaq and down Milne Inlet currently used during the average open water season (August 5 to October 15). BIM wants to increase production at the mine, which would necessitate more shipping to export the ore. The company has proposed starting to ship 2–3 weeks sooner in the summer (as of July 15), and later into the fall (until November 15; Bourbonnais et al., 2016). These shipping dates were proposed based on the analysis of CIS charts and satellite imagery (1980–2016) to understand the historical sea ice conditions in the region, and determine the vessel class, safety and feasibility of shipping in the shoulder seasons (Bourbonnais et al., 2016). The assessment concluded that shipping into the shoulder seasons was possible based on the use of various ice-strengthened vessel classes (Bourbonnais et al., 2016). Sikumiut are concerned about the impacts of icebreaking in the fall as tuvaq, along with the Tursukattak sinaa, are forming at this time, and changes to fall sea ice could impact travel safety throughout the subsequent winter and spring ice seasons. For example, shipping in the fall will leave large tracks of deformed, rough ice, dangerous for navigation during the dark months and cutting off traditional travel routes (**Figure 1**; Sikumiut, 2021). Sikumiut are also concerned that icebreaking earlier in the summer could further accelerate sea ice break-up and black carbon emissions from ships could change the albedo of the sea ice (Sikumiut, 2021). Changes to the sinaa and tuvaq could have critical consequences for Mittimatalingmiut for sea ice travel safety, in accessing hunting areas, for spring seal reproduction on the ice, and for polar bear migration. Additional concerns are due to the noise from icebreaking and the effects on local seal and narwhal populations (Sikumiut, 2021).

Discussions across many Sikumiut meetings evolved around the need to document Mittimatalik's historical sea ice conditions and develop a baseline of sea ice knowledge for the region. This sea ice baseline would be analyzed to understand:

- where and when the sea ice is changing to adapt sea ice travel; and

- how shipping later during sea ice freeze-up and earlier during sea ice break-up could compromise the safety of Mittimatalingmiut on-ice travel.

It was also important for Sikumiut to have this baseline to compare ongoing changes to sea ice, and the potential cumulative effects of shipping through the sea ice. To address Sikumiut's climate change adaptation and shipping impact questions, we needed to co-develop a novel way to create a Mittimatalik-specific sea ice climatology.

Available Date to Support Community-Scale Sea Ice Climatologies

Wilson started by reviewing the available satellite, CIS ice charts and *in-situ* datasets for the Mittimatalik region at a variety of scales to determine how additional data sources could supplement Sikumiut's IQ for a Mittimatalik specific sea ice climatology.

Satellite Data

The most widely used sea ice climatology comes from the Special Sensor Microwave Imager (SSM/I) satellites (NSIDC, 2021). SSM/I satellites have been imaging the polar regions since 1978, providing a 44-year-long database to monitor changing sea ice conditions (Stroeve and Meier, 2018). However, the spatial resolution of SSM/I imagery is on the order of 25 km, and community sea ice conditions are indiscernible from the topography of the Canadian Arctic archipelago in this imagery (Cooley et al., 2020; NSIDC, 2021).

Two other types of satellite sensors are optimal for sea ice monitoring: optical; and synthetic aperture radar (SAR). Optical satellites, such as NASA's Moderate Resolution Imaging Spectroradiometer (MODIS) and the European Space Agency's Sentinel-2 (ESA, 2019; NASA, 2019), are dependent on sunlight and therefore cannot image the earth's surface when there are clouds or during winter polar darkness in northern latitudes. MODIS images the Mittimatalik region daily at a resolution of 250 m and there is an archive of imagery dating back to the year 2000 (**Figure 1**). MODIS has been used successfully to develop climatologies of landfast ice break-up for Inuit communities using cloud free imagery during the spring and summer seasons with long daylight hours (Cooley et al., 2020). SARs, such as RADARSAT 1 and 2 (CSA, 2019) and Sentinel-1 (ESA, 2019), have their own energy source that send and receive microwave wavelengths to measure the roughness of the earth's surface. This built-in energy source allows for monitoring during the dark Arctic winters, approximately mid-November to mid-February (3 months). The microwave wavelengths of SARs can also penetrate most cloud cover, providing year-round imaging of the Arctic surface. The RADARSAT imagery archive dates back to 1997, with a majority of the imagery in a ScanSAR Wide beam mode with a 100 m resolution (**Figure 4B**).

Ice Charts

The longest recorded sea ice archive for Canada is based at the CIS (ECCC, 2021). Since 1968 the CIS has been monitoring sea ice to support summertime marine navigation and Arctic community re-supply (Shokr and Sinha, 2015). Between 1968 and 1995, detailed daily ice charts were produced using a combination of visual and SAR aerial reconnaissance missions, low-resolution satellite data, and meteorological information. In 1996, the CIS transitioned to using RADARSAT as their primary data source to operationally monitor sea ice in the Canadian Arctic (Ramsay et al., 1996, 1998). The CIS produces detailed daily ice charts for the major shipping routes in the Arctic during the summer season. In the fall, as the sea ice starts to freeze-up, ships leave the Arctic and the CIS transitions to weekly, less detailed regional charts to monitor the sea ice conditions over the winter months until break-up the following summer. The CIS archive now captures three 30-year climatological periods: 1971–2000; 1981–2010; and 1991–2020. CIS climatological products have been developed to generate sea ice climate normal maps and graphs to review change and variability in sea ice conditions in Canada. The CIS climatology has been created at a regional scale for the Western Arctic, Eastern Arctic and Hudson Bay and are not at a scale to capture the ice conditions for the Mittimatalik region (ECCC, 2021). However, the weekly charts in the CIS archive do provide some details of Mittimatalik ice conditions and are an additional data source for the community climatology (**Figure 4A**).

In-situ Observations

The Arctic Research Establishment (ARE) was a private research station run by the Steltner family based in Mittimatalik between 1975 and 1989. ARE collected oceanographic and sea ice data for ship engineering and ice-breaking research. Some Sikumiut members had worked for ARE taking measurements and requested that these data be relocated and returned to them. Between 2016 and 2018, Wilson searched Canadian archives, contacted retired scientists and eventually connected with members of the Steltner family. The data collection had been kept in the family home in southern Ontario and the data included environmental observations recorded in field books, reports, photographs, and on film. Bell sought funding to archive this dataset, and between 2019 and 2020 the collection was scanned, sorted, and boxed up. Digital copies of the collection are now in the community of Mittimatalik, but several years of work are still required to review and enter the observations into a database for research. The Steltner family donated the ARE collection to the Government of Nunavut and the physical records are now stored in their archives currently housed in Ottawa.

Community-based monitoring (CBM) has been gaining significant interest to fill gaps in sparse Arctic environmental information (Johnson et al., 2015). The benefits of CBM approaches include year round monitoring, conducted by the Indigenous peoples who live in the region, and in providing local scale information that Arctic communities can use to address their own research needs (Johnson et al., 2015). SmartICE is a CBM service that was established to monitor sea ice in the community of Mittimatalik in 2016. However, the current length of the SmartICE record (5 years) is not yet long enough for use in the Mittimatalik climatology.

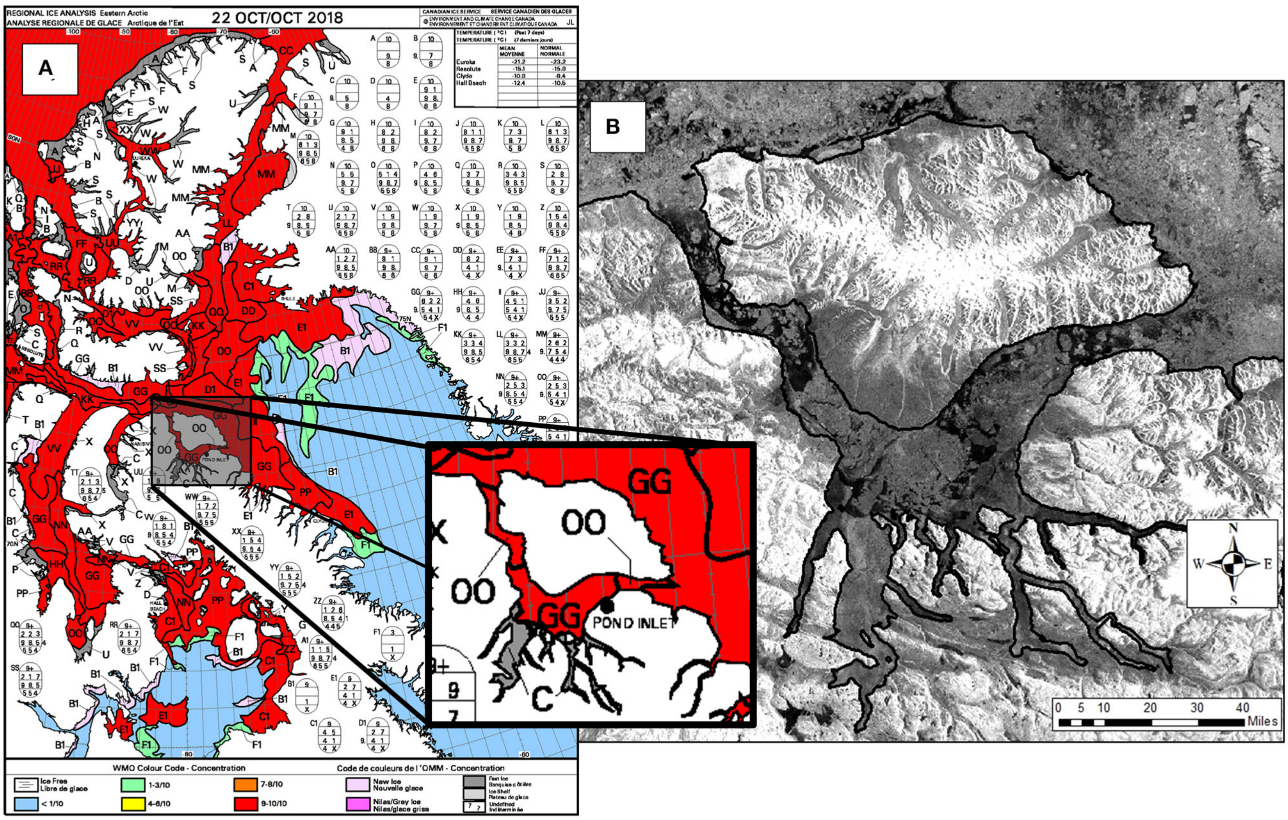

FIGURE 4 | (A) CIS eastern Arctic regional ice chart for October 22, 2018 (ECCC, 2020); **(B)** RADARSAT-2 ScanSAR Wide image of the Mittimatalik region, October 22, 2018 (CSA, 2019).

Inuit Qaujimajatuqangit

Inuit hold the only long term and consistent record of sea ice in the Canadian Arctic. Riedlinger and Berkes (2001) discuss how IQ is a source of climate history and can provide a baseline to assess change and fill Arctic monitoring gaps. However, in reviewing the literature we found no practical examples where IQ was mobilized for its climate history.

The Sikumiut maps that were co-produced in 2018 share the IQ of known locations of safe and hazardous ice conditions by season (**Figures 2A**, **3A,B**). The winter travel map highlights dangerous areas such as reoccurring *naggutiit* (cracks in the ice that can be easily crossed), *ivujuk* (ridges, high areas of rough ice you have to travel around), and *siku saattuq aragulimaamik* (thin ice all year; **Figure 3A**). The spring maps show new and expanding dangerous travel areas such as *aajurait* (leads, cracks in the sea ice that get wider in the spring are always possible to cross), *siku saattuq upingaat pigiarningani* (thin ice in spring), and *imaqainnaujattuq ukiutamaa* (water that runs from the glaciers; **Figure 3B**). These maps provide an IQ-based climatology for the region of Mittimatalik; however, the information on which they are based is not in a database, they exist in the collective memory of Sikumiut members.

Based on the assessment of available sea ice information sources for Mittimatalik we had the following four: (1) Sikumiut's IQ; (2) the CIS charts (1968 to present); (3) RADARSAT 1 and 2 (1997 to present) imagery; and (4) MODIS imagery (2000 to present). The overlapping time period of the available information was from 1997 to 2019, a 23-year time period, slightly less than a standard 30-year climatology. We then began to explore how IQ could interpret and review the satellite and ice chart data to develop a Mittimatalik specific sea ice climatology based on IQ.

METHODS

The co-production of the Mittimatalik sea ice atlas occurred over 3 years between 2019 and 2021, as outlined in **Table 2**. During 2019 a majority of the co-development and training was done in person in Mittimatalik. As the COVID-19 pandemic hit and travel restrictions were implemented, we continued our collaborative work by mailing data to each other on external drives and moving our training, discussions and meetings on-line (**Table 2**). Bandwidth limitations in the community reduced the use of videoconferencing as a collaboration platform, and a majority of our interactions were by text, telephone and e-mail in 2020 and 2021. This section illustrates our preliminary steps, the development and analysis of the break-up and freeze-up maps, and the

TABLE 2 | Mittimatalik sea ice atlas co-production timelines and responsibilities.

<table>
<tr><th colspan="2">Year</th><th>Month</th><th>Arreak</th><th>Wilson</th></tr>
<tr><td colspan="2">2018</td><td>June–December</td><td></td><td>Archiving satellite imagery and CIS charts.</td></tr>
<tr><td colspan="2" rowspan="8">2019</td><td>February</td><td colspan="2">In Mittimatalik: initial discussion on methods to interpret and map break-up.</td></tr>
<tr><td>March</td><td></td><td>Develop remote sensing training.</td></tr>
<tr><td>April</td><td colspan="2">In Mittimatalik: Remote sensing training.</td></tr>
<tr><td>May–July</td><td>Remote sensing interpretation practice: monitoring spring break-up conditions with satellite imagery on the SIKU website.</td><td>Develop training procedures for satellite imagery analysis and digitizing break-up.</td></tr>
<tr><td>July</td><td colspan="2">In Mittimatalik: external drive with archived satellite imagery provided to Arreak; training on interpretation and digitization of archived break-up imagery, and discussion on methods to interpret and map freeze-up. Start of satellite imagery analysis for break-up.</td></tr>
<tr><td>September</td><td colspan="2">In Mittimatalik: reviewing work, sorting out issues.</td></tr>
<tr><td>October</td><td rowspan="2">Satellite imagery analysis and digitization for break-up continued.</td><td>Freeze-up data pre-processing: converted CIS charts to raster, extracted ice type and fast ice parameters.</td></tr>
<tr><td>December</td><td>Develop training procedures for freeze-up analysis: creating weekly average maps and yearly freeze-up maps in ArcMAP and graphing trends in Excel.</td></tr>
<tr><td rowspan="7">2020</td><td rowspan="13">COVID 19 Pandemic</td><td>February</td><td colspan="2">In Mittimatalik: Training on freeze-up analysis of CIS charts.
External drive with freeze-up raster files provided to Arreak. Last trip before COVID.</td></tr>
<tr><td>March–May</td><td>Break-up GIS files copied to back-up external drive and mailed to Wilson.</td><td>Break-up data processing: Converting digitized weekly maps to raster for analysis.</td></tr>
<tr><td>August</td><td>Freeze-up analysis: developing weekly average maps, yearly freeze-up maps, and graphing trends.</td><td>Develop training procedures for break-up analysis: create weekly average maps, yearly freeze-up maps, and graphing trends. Mailed copy of break-up raster files and analysis procedures on external drive to Arreak.</td></tr>
<tr><td>September</td><td colspan="2">E-mail freeze-up maps and graphs to Wilson.
Review freeze-up analysis and discuss results by phone.</td></tr>
<tr><td>October</td><td>Break-up analysis: Produce weekly average maps, yearly freeze-up maps and trend graphs.</td><td>Testing initial color schemes and legends.</td></tr>
<tr><td>November</td><td colspan="2">E-mail maps and graphs to Wilson.
Review break-up analysis and discuss results by phone.</td></tr>
<tr><td>December</td><td colspan="2">Sikumiut meeting: Initial results presented by Arreak (Wilson and Bell by phone).</td></tr>
<tr><td rowspan="6">2021</td><td>January</td><td colspan="2">Finalizing map color schemes for visual accessibility and printing.</td></tr>
<tr><td>March</td><td colspan="2">Draft #1 of freeze-up maps printed and mailed to Mittimatalik. Sikumiut meeting to review draft freeze-up maps (Wilson and Bell by phone).</td></tr>
<tr><td>May</td><td colspan="2">Draft #1 break-up maps and draft #2 freeze-up maps and text printed and mailed to Mittimatalik. Sikumiut meeting: review of draft maps and translated text (Wilson and Bell by phone). Revisions to maps.</td></tr>
<tr><td>June–August</td><td colspan="2">Layout, drafting text and translation into Inuktitut.</td></tr>
<tr><td>September–October</td><td colspan="2">Review of draft atlas with translated text, make revisions.</td></tr>
<tr><td>November–December</td><td colspan="2">Printing of atlas and shipping to Mittimatalik.</td></tr>
</table>

process to create maps that were accessible and intuitive for Mittimatalingmiut.

Preliminary Work

In 2018 Wilson began visually reviewing and archiving RADARSAT-1 (1997–2013) and RADARSAT-2 (2009–2019) imagery between October and July. Cloud free MODIS (2000–2019) imagery were visually reviewed between mid-February to the end of October when the region has adequate daylight hours for optical imagery (NASA, 2019). Weekly satellite coverage of the Mittimatalik area averaged 3 per week with RADARSAT data and an additional 2 per week with MODIS data during the freeze-up and break-up periods, totalling ∼4,000 images archived. Additionally 500 CIS weekly charts were also archived from the CIS (ECCC, 2021).

Once the data was archived, we began planning training for Arreak to learn how to interpret the satellite imagery. Optical imagery is fairly easy to interpret because it is very similar to a color photograph. However, SAR imagery can be difficult to interpret for untrained users and requires a shift in thinking to understand that these images represent the surface roughness of the earth. For example, dark smooth areas in SAR imagery can commonly be areas of open water and/or smooth sea ice. The goals of this pilot satellite imagery training were two-fold: (1) so Arreak could interpret the satellite imagery using his IQ to map the safe and unsafe sea ice travel conditions around Mittimatalik from 1997 to 2019; and (2) so SmartICE operators could start using publicly available satellite data from SIKU and Polar View on-line platforms in their day-to-day SmartICE operations (Polar View, 2019; Arctic Eider Society, 2020).

In April 2019, a 4-day satellite interpretation training session was held in Mittimatalik to pilot this training with Arreak and two other Inuit SmartICE operators from Qikiqtarjuaq (Jenny Mosesie) and Arviat (Robert Karetak) (Wilson et al., 2020). This training was then put into practice between May and July with the three SmartICE operators monitoring their regions in near real-time during the 2019 sea ice break-up season by accessing the satellite imagery on the SIKU website (**Table 2**).

Break-Up Maps

Arreak and Wilson began co-developing the IQ-based sea ice climatology methods in February 2019 (**Table 2**). We began by looking at the spring and early summer satellite imagery together to understand what sea ice features could be identified in the imagery, and what was important from an Inuit perspective to capture in the imagery.

The interpretation of sea ice in satellite imagery for charting is based on an international standard established by the World Meteorological Organization (WMO). The Manual of Standard Procedures for Observing and Reporting Ice Conditions (MANICE) defines and describes the navigational terms for sea ice (ECCC, 2005). The MANICE terms evolved primarily by identifying sea ice from a bird's eye view using aircraft and helicopters from the 1960's to 1990's, and since the late 1990's using predominantly satellites. We reviewed Sikumiut's sea ice terms to determine if we could use Inuit specific ice types instead of the MANICE ice types to classify the satellite imagery. It was difficult to identify these specific ice types during break-up at the resolution of the MODIS (250 m) and RADARSAT ScanSAR Wide (100 m) imagery. While the MANICE terms evolved from above looking down at the sea ice surface, the Inuktitut sea ice terms evolved from traveling on the sea ice, at a scale of <1 m (Wilson et al., in press). The spatial scales of the Sikumiut sea ice terms did not align with the scale of the available satellite imagery. We then discussed classifying the imagery using the MANICE sea ice types since they were at the scale of the satellite imagery, however for break-up the MANICE types do not indicate the stage of melt or break-up. For example, ice that is classified as thick first year ice in May, will remain this ice type until the area completely melts and becomes open water.

Ice charts describe sea ice conditions using a numeric code called "the egg code" (ECCC, 2005). Numbers are used in the egg code to eliminate language barriers in the polar navigational community. Polygons are drawn on the satellite imagery around homogenous areas of sea ice and the numeric egg code describes up to three sea ice types, their concentrations (expressed in tenths) and floe sizes within the polygon (**Figure 4A**). Using these eggs codes, captains navigate through ice-free, or lower concentrations of ice, avoiding higher concentrations of moving ice dangerous for navigation. Estimating sea ice concentrations for the Mittimatalik climatology was also discussed. For example, break-up is often based on when ice concentrations, are <5/10ths (Archer et al., 2017; Segal et al., 2020b). Arreak did not feel that 5/10th concentration was a useful threshold to determine break-up in Mittimatalik. Break-up in the area does not occur all at once, it occurs in different areas and at different times, and is often linked to the stability of the sinaangit.

What Arreak felt was climatologically important to map were locations of rough sea ice, aajurait, sinaagnit, and areas of sea ice breakup that were no longer safe for travel (open water and/or areas with numerous breaks in the ice). We first looked at roughness, as SAR imagery has been used to develop sea ice surface roughness maps for Inuit travel (Segal et al., 2020a). However, when traveling on the ice, areas in the SAR image that are rough can actually be smooth for sea ice travel with sufficient snow cover. In the spring, as puddles and melt ponds form on the sea ice, the presence of water dominates the SAR backscatter resulting in smooth areas on the SAR image, masking the ice surface underneath. For the purpose of this historical analysis, we were concerned that ice roughness would be overestimated in winter and underestimated during spring melt. Therefore, we removed sea ice roughness as a parameter and focused on mapping aajurait, sinaagnit, and areas of break-up. The latter were defined as areas that were no longer safe for travel. The break-up areas could include open water, melting sea ice and/or areas with multiple aajurait, which would no longer be safe to travel on.

Wilson used the CIS climatology methods as initial inspiration for the Mittimatalik climatology. Using the same climatological weeks as the CIS, Arreak reviewed and interpreted the satellite data for each week. Arreak was trained using ArcMap 10.5 Geographic Information Systems (GIS) software to digitize the weekly locations of aajurait, sinaagnit and areas of break-up. Arreak spent half of his time over 6 months (**Table 2**) interpreting the imagery and digitizing maps. Arreak interpreted each week of the archived satellite data from late May until early August to create 10 weekly maps per year. This weekly analysis for break-up was repeated for each year from 1997 to 2019 (23 years), to create 230 weekly maps, analyzing ∼2,000 satellite images in total.

As Arreak and Wilson reviewed the satellite data, they made notes detailing:

- the dates when the snow melted, and when the sea ice became visible in the MODIS imagery;
- when areas of open water on the sea ice first became visible in the MODIS and RADARSAT imagery; and
- the final break-up dates for the Tursukattak and Navy Board sinaangit as detected in the MODIS and/or RADARSAT imagery (±2 days).

The RADARSAT SCW data was block averaged to reduce speckle for interpretation, reducing the resolution to 200 m. The MODIS imagery was interpreted with a resolution of 250 m. Wilson converted the weekly break-up polygons to raster in ArcMAP with a cell size of 500 m^2. Each cell in the maps were assigned a value of 1 for break-up and 0 for tuvaq. Training focused on ArcMap spatial analysis tools to create weekly and yearly maps of average ice conditions, and to compare differences between years. Arreak developed weekly average break-up maps by adding together all the maps for the same climatological week over the 23-year record (1997–2019). The summed values provided an indication of how often break-up occurred in this cell over the 23-year record. For example, if the summed value was equal to 18, this meant that break-up occurred in this particular cell 18 times out of 23 years, or 78% of the time. The categories in the weekly maps were developed to indicate the following safe travel conditions: (1) dangerous; (2) frequently dangerous; (3) sometimes dangerous; and (4) generally safe (**Table 3**). The total area of break-up was calculated to determine and compare how much of the Mittimatalik region was breaking up each week. These percentages were exported to Microsoft Excel and Arreak generated graphs to analyse trends and variability in Mittimatalik's sea ice conditions over 23 years. Wilson performed linear regressions and tested the regressions for statistical significance.

Sikumiut had mentioned on several occasions that the greatest change in sea ice has occurred in the last decade. While graphs can indicate trends and variability in break-up over the years, we wanted to develop maps to understand where break-up was occurring earlier. Using the same procedures for the weekly frequency of break-up maps, Wilson summed the maps for the same climatological weeks for the first 13 years (1997–2009) and the last 10 years (2010–2019). These maps were reclassified into four categories based on how often break-up was occurring in the area in the two separate time periods: 0–25% of the time; 25–50% of the time; 50–75% of the time; and 75–100% of the time (**Table 4**). The two reclassified time period maps were then added together to produce unique cell values that were grouped into 5 categories to indicate where break-up has changed the most during the last 10 years: earlier; sometimes earlier; no change; sometimes later; and later (**Table 4**).

Freeze-Up Maps

Post-analysis of sea ice freeze-up in the MODIS and RADARSAT satellite imagery proved challenging. It was difficult to historically map the fluid and dynamic sea ice conditions that moved with the winds and ocean currents until they consolidate in early winter (**Figure 4B**). We again looked to the weekly CIS charts, as they were created using satellite data and meteorological observations in near-real time (**Figure 4A**). We discussed using the ice charts concentrations as a way to classify freeze-up, based on a threshold of concentrations >5/10ths (Archer et al., 2017; Segal et al., 2020b). Again, what Arreak felt was most important to know during freeze-up was when the sea ice was safe to travel on, and when the sinaagnit were forming, the 5/10ths threshold did not convey this information. We also looked at the MANICE ice types to infer the thickness of the sea ice. For example, estimating ice types >1 foot (30 cm) as safe for travel. While some hunters are experienced and knowledgeable to travel on newer ice types, for most community members safe travel is considered possible once the ice becomes tuvaq (Wilson et al., in press).

The CIS charts do code tuvaq once first-year ice concentrations reach 9+ and 10/10ths (**Figure 4A**). As a result, we used the CIS weekly ice charts over a 13-week period between October and December to capture Mittimatalik freeze-up. Historically, ice chart production ceased for the Mittimatalik region near the end of November as the sea ice froze and ships left the region, therefore there are no weekly ice charts available for the month of December between 1997 and 2005. With improved satellite coverage starting in 2006, the CIS began producing weekly charts into the winter months. Benoit Montpetit (ECCC Wildlife S&T Branch) developed scripts for us to extract the landfast ice polygons from the charts and convert to raster. Each cell in the maps were assigned a value of 1 for tuvaq and 0 if it wasn't tuvaq. The production of freeze-up average weekly maps, difference maps, yearly maps and trends and variability analysis followed the same steps as for break-up.

Accessible Atlas Colors and Legends

As the siku asijjipallianinga was going to be something completely new for Mittimatalingmiut, it was important to develop maps that were intuitive, culturally accessible, and distinct by season and map type. We spent several months testing different color schemes for the maps in the atlas. Certain colors tend to be intuitive, for example green for safe, red for dangerous and blue for water. Red and green diverging colors were not used in the same map out of considerations for people with color blindness. Red and blue, pink and green, and purple and orange are recommended contrasting colors for color accessibility (Brewer et al., 2002). We tested using red for dangerous conditions and blue for safer conditions in the weekly average travel freeze-up maps. However, for Inuit, dangerous sea ice travel conditions are often because of open water, so using blue to indicate safer travel conditions was counter intuitive. We reached consensus on using the contrasting colors of green to indicate safer travel conditions and pink for more dangerous travel conditions for the weekly average travel maps.

With 6 different maps in the atlas we were concerned that having 6 different legends would be confusing for users. For the weekly average travel maps, we tested and refined using green for safer travel conditions and pink for more dangerous travel conditions in order to have the same color scheme for freeze-up and break-up (**Table 5**). The categories in the weekly maps were also developed so they could be used in both the freeze-up and break up maps (**Tables 3**, **5**). For the difference maps, we also tested a color scheme that could be used for both the freeze-up and break-up. Orange to indicate earlier freeze-up or break-up, and purple to indicate later freeze-up or break-up (**Table 5**). Once again, the categories in the difference maps could be used for both freeze-up and break-up: (1) earlier; (2) sometimes earlier; (3) no change; (4) sometimes later; and (5) later (**Table 5**).

For the yearly maps, a sequential color scheme was more intuitive and preferred by all. For enough contrast in viewing and printing sequentially colored maps, no more than 6 shades of the same color are recommended (Brewer et al., 2002). We selected a red sequential color scheme for break-up so red could indicate dangerous travel areas (**Table 5**). Arreak initially digitized 10 weeks for break-up, but in the end, we found that negligible break-up occurred in the first 3 weeks (May 28 to June 27) of the record, so these 3 weeks were removed. In the end, yearly break-up maps in the atlas represent 7 weeks, between June 18 and August 5; from 1997 to 2019 (**Table 5**). We could not reduce

TABLE 3 | Weekly average break-up map categories.

Weekly frequency of break-up 1997–2019 (23 years total)			Average travel conditions
# of years the area was breaking-up	Percentage of time the area was breaking-up	Reclassified value	
1–5 years	0–25%	1	Generally safe
6–10 years	25–50%	2	Sometimes dangerous
11–16 years	50–75%	3	Frequently dangerous
17–23 years	75–100%	4	Dangerous

TABLE 4 | Classifications for the difference in the frequency of break-up maps for two time periods.

1997–2009 First 13 years			2010–2019 Last 10 years			Difference map First 13 + Last 10 values			
# of years	% of time	Reclassified value	# of years	% of time	Reclassified value	New value		Percent change	Legend category
0–4	0–25%	0	0–2	0–50%	2	0 + 2 =	2	0	No change
0–4	0–25%	0	3–5	25–50%	20	0 + 20 =	20	+25%	Sometimes earlier
0–4	0–25%	0	6–7	50–75%	200	0 + 200 =	200	+50%	Earlier
0–4	0–25%	0	8–10	75–100%	2,000	0 + 2,000 =	2,000	+75%	Earlier
5–7	25–50%	−10	0–2	0–50%	2	(−10) + 2 =	(−8)	(−25%)	Sometimes later
5–7	25–50%	−10	3–5	25–50%	20	(−10) + 20 =	10	0	No change
5–7	25–50%	−10	6–7	50–75%	200	(−10) + 200 =	190	+25%	Sometimes earlier
5–7	25–50%	−10	8–10	75–100%	2,000	(−10) + 2,000 =	1,990	+50%	Earlier
7–9	50–75%	−100	0–2	0–50%	2	(−100) + 2 =	(−98)	(−50%)	Later
7–9	50–75%	−100	3–5	25–50%	20	(−100) + 20 =	(−80)	(−25%)	Sometimes later
7–9	50–75%	−100	6–7	50–75%	200	(−100) + 200 =	100	0	No change
7–9	50–75%	−100	8–10	75–100%	2,000	(−100) + 2,000 =	1,900	+25%	Sometimes earlier
10–13	75–100%	−1,000	0–2	0–50%	2	(−1,000) + 2 =	(−998)	(−75%)	Later
10–13	75–100%	−1,000	3–5	25–50%	20	(−1,000) + 20 =	(−980)	(−50%)	Later
10–13	75–100%	−1,000	6–7	50–75%	200	(−1,000) + 200 =	(−800)	(−25%)	Sometimes later
10–13	75–100%	−1,000	8–10	75–100%	2,000	(−1,000) + 2,000 =	1,000	0	No change

the number of weeks to six to meet printing recommendations, but in reviewing the printed maps, we felt there was sufficient contrast for the 7 weeks.

The yearly freeze-up maps initially showed freeze-up over 13 weeks, too many classes for a single color scheme. Negligible freeze-up occurred between October 1 and 21 over the record, so these 3 weeks were removed. Very little change in freeze-up also occurred during the following 2-week periods of (1) October 22 to November 4 when freeze-up is just starting; (2) December 4–16; and (3) December 17–20 when the sea ice growth slows as it consolidates. These three, 2-week periods were merged reducing the number of classes for the yearly freeze-up maps to eight (**Table 5**). A sequential three-color scheme used yellow for late October; green for November; blue for December; and dark blue for remaining areas of open water at the end of December (**Table 5**; Brychtová et al., 2015). For the freeze-up and break-up yearly maps, the lightest colors indicate the areas in which sea ice is present for the longest period of time and the darkest colors where sea ice was present for the shortest amount of time.

TABLE 5 | Mittimatalik siku asijjipallianinga legend categories and color schemes.

<table>
<tr><th>Atlas maps</th><th colspan="8">Legend color/category</th></tr>
<tr><td rowspan="4">1. Weekly average travel conditions for freeze-up and break-up</td><td colspan="8">Dangerous</td></tr>
<tr><td colspan="8">Frequently dangerous</td></tr>
<tr><td colspan="8">Sometimes dangerous</td></tr>
<tr><td colspan="8">Generally safe</td></tr>
<tr><td rowspan="5">2. Weekly difference maps for freeze-up and break-up</td><td colspan="8">Earlier</td></tr>
<tr><td colspan="8">Sometimes earlier</td></tr>
<tr><td colspan="8">No change</td></tr>
<tr><td colspan="8">Sometimes later</td></tr>
<tr><td colspan="8">Later</td></tr>
<tr><td>3. Yearly freeze-up maps</td><td>Oct 22 to Nov 4 2 weeks</td><td>Nov 5–11</td><td>Nov 12–18</td><td>Nov 19–25</td><td>Nov 26 to Dec 2</td><td>Dec 3–16 2 weeks</td><td>Dec 17–30 2 weeks</td><td>Open water</td></tr>
<tr><td>4. Yearly break-up maps</td><td>June 11–24</td><td>June 25 to July 1</td><td>July 2–8</td><td>July 9–15</td><td>July 19–22</td><td>July 23–29</td><td colspan="2">July 30 to Aug 5</td></tr>
<tr><td rowspan="2">Across all maps</td><td colspan="8">Outside travel region</td></tr>
<tr><td colspan="8">Land</td></tr>
</table>

Finally, we also wanted to ensure that each color was used only once for consistency across all the maps, for example not using blue for ice in one map and blue for water in another map. Although not perfect, considerable effort was put into selecting the colors and developing the legends to reduce the number of legends from 6 to 4 and to ensure they were accessible and culturally intuitive for Mittimatalingmiut (**Table 5**). Sikumiut reviewed the maps and legends at meetings in December 2020, March and May 2021 (**Figure 3B**). During these meetings we also discussed what we would call this sea ice climatology in Inuktitut. Sikumiut decided on the "Mittimatalik siku asijjipallianinga" (changes of the sea ice).

RESULTS

The Mittimatalik siku asijjipallianinga project includes the following 14 products to capture the sea ice climatology for the community between 1997 and 2019. Samples of these products are illustrated below (**Figures 4–15**) as we review the averages, trends, and variability in the sea ice freeze-up and break-up seasons over the 23-year climatology.

Freeze-up, October 22 to Dec 20, (1997–2019):

1) Ten weekly average tuvaq maps (e.g., **Figure 5**)
2) Summary graph of average tuvaq formation by week (**Figure 6**)
3) Summary graph showing the weekly variability in tuvaq formation (**Figure 7A**)
4) Summary graph illustrating the weekly frequency of tuvaq formation (**Figure 7B**)
5) Twenty-three maps showing the spatial formation of tuvaq for each year (e.g., **Figure 8**)
6) Six weekly difference maps showing areas where tuvaq is forming earlier or later in the last 10 years (e.g., **Figure 9**)

Break-up, June 18 to July 29, (1997–2019):

7) Frequency graph illustrating the key indicators for break-up (**Figure 10A**)
8) Graphs of the Navy Board and Tursukattak sinaangit average break-up dates (**Figures 10B,C**)
9) Six weekly average break-up maps (e.g., **Figure 11**)
10) Summary graph of average break-up by week (**Figure 12**)
11) Summary graph highlighting the weekly variability in break-up (**Figure 13A**)
12) Summary graph illustrating the critical weeks for break-up (**Figure 13B**)
13) Twenty-three maps showing spatial break-up of sea ice for each year (e.g., **Figure 14**)
14) Six weekly difference maps showing areas where the sea ice is breaking up earlier or later in the last 10 years (e.g., **Figure 15**).

Freeze-Up Results

For the week of November 5–11, there is an average of 38% (std dev 35%) tuvaq in the region with initial areas of tuvaq forming in the southern inlets and sounds; however, the sea ice is not normally safe for community travel (**Figure 5A**). By the weeks of November 12–18 and 19–25, tuvaq formation averages 58–71% (std dev 35–32%), both sinaangit are establishing in Navy Board and Tursukattak, and normally the sea ice is safe for Mittimatalingmiut to travel in the southern inlets and sounds (**Figures 5B,C**). While the sea ice in Navy Board Inlet is generally safe for travel on by November 19–25, it is normally inaccessible until the formation of tuvaq in Tasiujaq. On average, tuvaq increases to 80% (std dev 27%) during the week of November 26-December 3 and Mittimatalingmiut are normally able to travel from the community west into Tasiujaq (**Figure 5D**). By the week of December 24–30, the region averages 97% (std dev 4%) tuvaq and Mittimatalingmiut are normally traveling to the Tursukattak sinaa (**Figure 5E**).

While freeze-up may be occurring later in other areas of the Arctic, we found no significant trends in the weekly formation of tuvaq between 1997 and 2019. These negligible trends are a result of the high variability in the formation of tuvaq during freeze-up between 1997 and 2019 (**Figure 6**). However, this

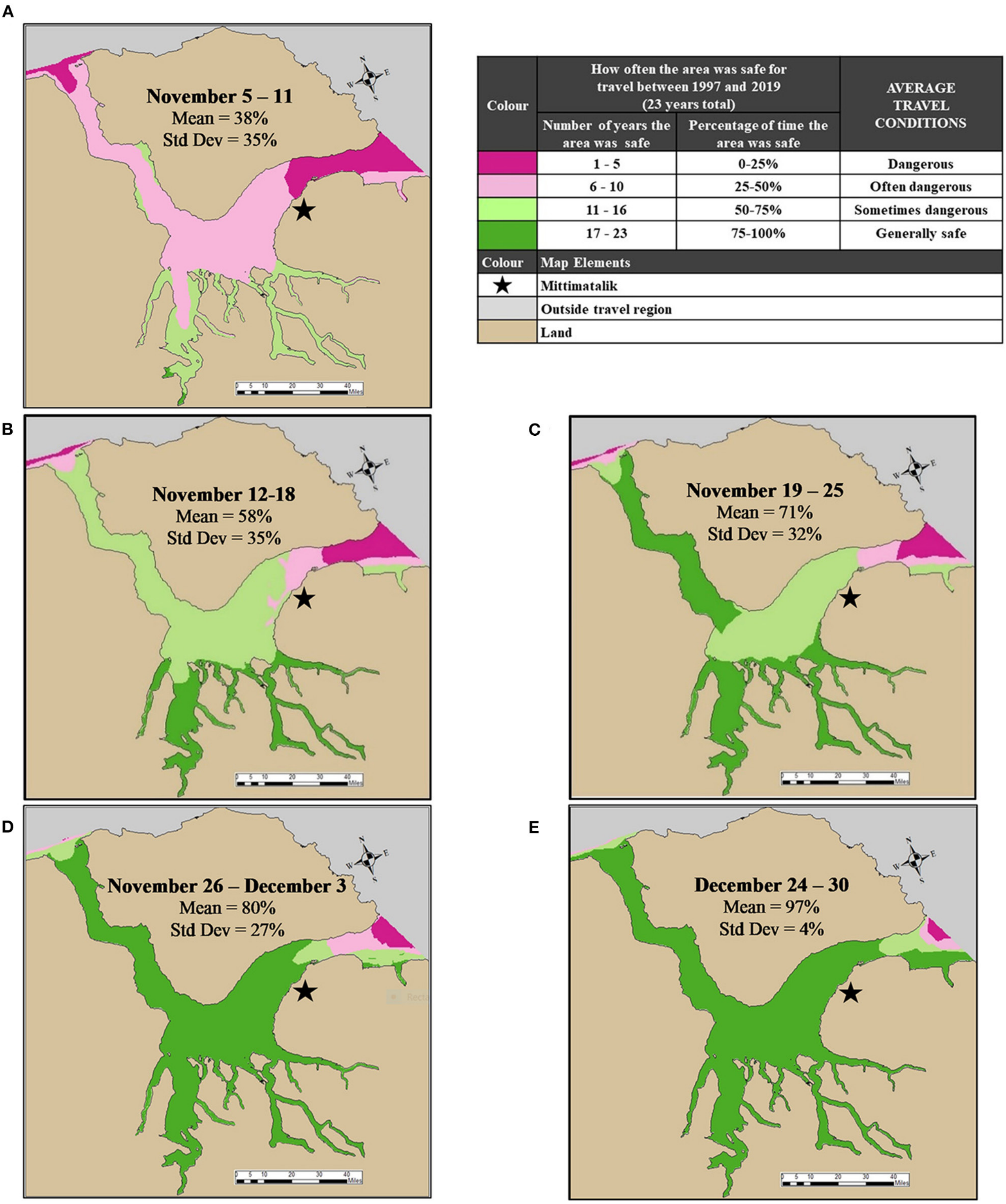

FIGURE 5 | Weekly average tuvaq maps showing freeze-up for the Mittimatalik region, 1997–2019. **(A)** Average tuvaq, November 5–11. **(B)** Average tuvaq, November 12–18. **(C)** Average tuvaq, November 19–25. **(D)** Average tuvaq, November 26–December 3. **(E)** Average tuvaq, December 24–30.

variability is high only for particular weeks during freeze-up. The initial freeze-up week of October 29 to November 4 shows moderate variability, with an inter-quartile range (IQR) of 21% (**Figure 7A**). The outliers correspond to the years of 2002 and 2018 that had unusually high percentages of tuvaq early in the ice season (80 and 92%, respectively; see **Figure 8A** for the 2018

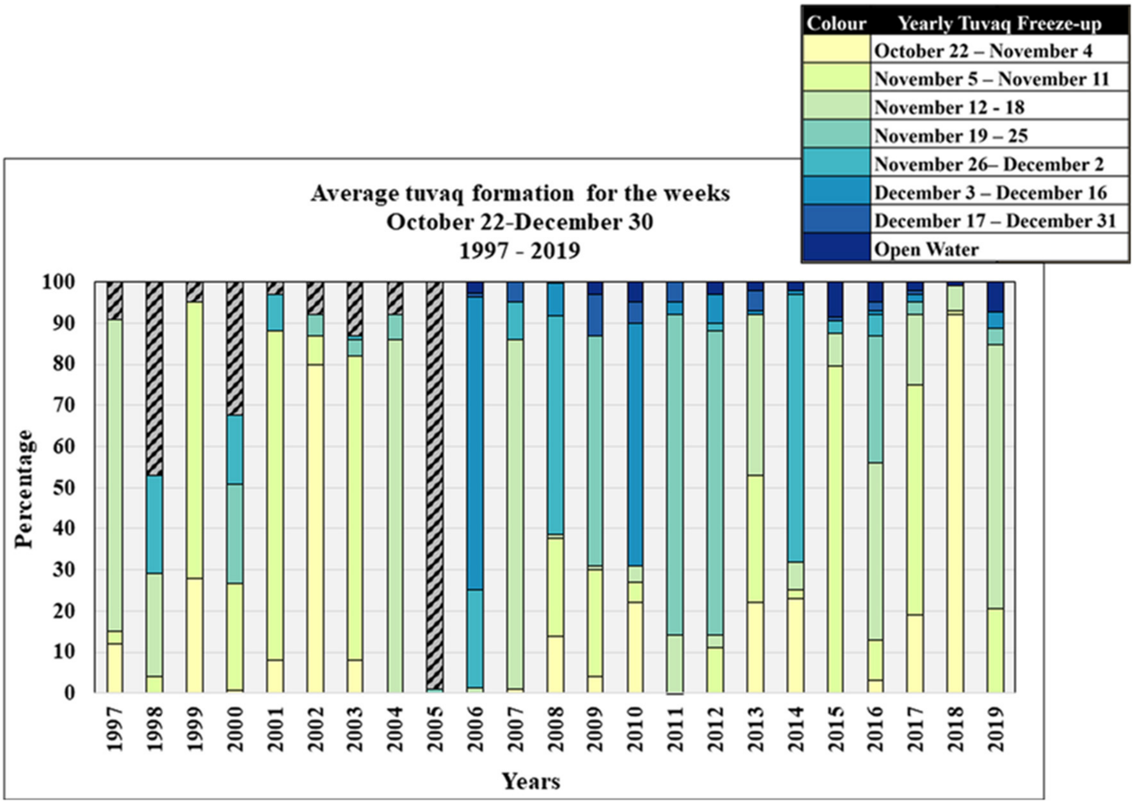

FIGURE 6 | Summary graph of average tuvaq formation for freeze-up, 1997–2019. Each bar is a year showing the weekly percentage of tuvaq freeze-up by color: yellow for late October; green for November; blue for December; and dark blue for remaining areas of open water at the end of December. Years with more blue represent the late formation of tuvaq. Years with more yellow represent the early formation of tuvaq.

map). The subsequent 3 weeks show the largest variability in tuvaq formation: November 5–11 with an IQR of 70%; November 12–18 with an IQR of 58%; and November 19–25 with an IQR of 47%. Later into the freeze-up season, this variability decreases significantly with an IQR of 4–7% for the weeks of November 26 to December 2, December 3–16, and December 17–20. The week of November 26 to December 2 had five outlier years corresponding to 1998, 2000, 2005, 2006, and 2010, in which tuvaq formation was unusually late. The 2005 freeze-up season had only 1% tuvaq by this week and the 2006 season had the second lowest percentage of tuvaq at 25% (see **Figure 8B** for the 2006 map).

A visual analysis of the yearly tuvaq freeze-up maps showed no spatial differences in where the tuvaq and sinaangit formed initially, or their subsequent expansion in early, average or late freeze-up years. While there is large variability for when the sea ice freezes, the spatial patterns for progressive expansion of tuvaq and sinaangit were highly consistent throughout the climatology. The weekly average maps (**Figure 5**) capture this consistent spatial pattern of freeze-up for all years except 1998 when tuvaq formed last in Tasiujaq (**Figure 5**; see **Figure 8C** for 1998 map).

To understand which weeks were critical for tuvaq formation during freeze-up, those with the highest percentages of tuvaq formation were tabulated for each year from 1997 to 2019 (**Figure 7B**). The weeks with the highest frequency of tuvaq formation were November 5–11 (26%) and November 12–18 (30%). Together, these 2 weeks comprise on average 56% of the annual formation of tuvaq and highlight the importance of this freeze-up period in early November.

The weekly difference maps show the spatial change in tuvaq within the last 10 years (**Figure 9**). The week of November 5–11 shows that tuvaq is forming earlier in some of the southern inlets and sounds (**Figure 9A**). The weeks of November 19 to December 2 show that tuvaq has been freezing up earlier in Tasiujaq and into Navy Board Inlet (**Figures 9D,E**). These results are counter intuitive to our expectations. Because we are mapping immobile tuvaq, this earlier freeze-up cannot be due to an increase of imported ice. Sikumiut were also perplexed to see freeze-up happening earlier in certain areas and during certain weeks, as this does not align with their IQ. It would be interesting to have Inuit map the freeze-up of sea ice in real-time to compare with the CIS charts to understand if there are differences in how Inuit and the CIS would interpret tuvaq freeze-up.

Break-Up Results

The start of the break-up season begins with snowmelt on land. Snowmelt increases local river runoff, flooding and melting the

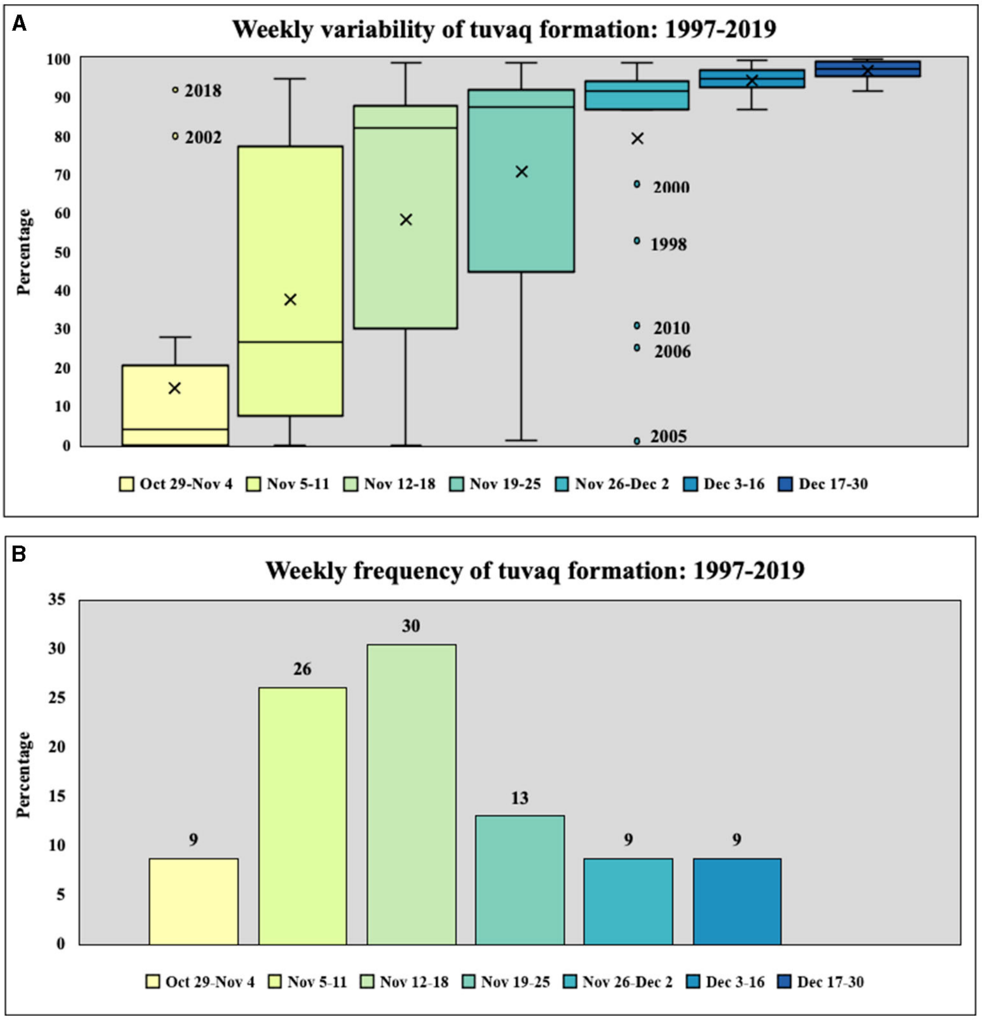

FIGURE 7 | (A) Summary graph of the weekly variability in tuvaq formation for freeze-up, 1997–2019. The box outlines the interquartile range, the average range in the variability of tuvaq formation for a particular week over the 23-year period (1997–2019). The line through the box is the median and the X denotes the mean. The vertical "whisker" lines show the minimum and maximum values. The dots correspond to outliers, or years with unusual tuvaq percentages. **(B)** Weekly frequency of tuvaq formation, 1997–2019.

sea ice at the mouths of rivers. The onset of snowmelt was detectable in the MODIS imagery in 17 of 23 years (74%) for the week of June 11–17 (**Figure 10A**). By the following week of June 18–24, areas of open water became visible in the satellite imagery in the southeast inlets and mouths of local rivers, as was captured in the average break-up maps (**Figure 11A**). Typically, the sea ice is still safe for travel during the week of June 25 to July 1 with an average of only 7% (std dev 7%) of the area breaking-up (**Figure 11B**). By July 2–8, the area is averaging 19% (std dev 13%) break-up. Areas that are no longer safe for sea ice travel are expanding in the south and southeast sounds and inlets, and along the coastlines. Travel to both sinaangit are less safe (**Figure 11C**). The week of July 9–15 shows how quickly the break-up season advances (**Figure 11D**). While the region on average is 47% (std dev 24%) broken-up, break-up around the community is advanced, and Mittimatalingmiut are no longer able to access safe areas for sea ice travel from the community. By July 16–22 the area averages 80% (std dev 21%) break-up (**Figure 11F**) and the Tursukattak and Navy Board sinaangit normally break-up this week (**Figures 10B,C**). On average, by the week of July 23–29 the area is 94% (std dev 8%) broken-up (not shown), and Mittimatalingmiut are waiting for the remaining ice to melt, or be exported by winds and ocean currents, to begin hunting and fishing by boat.

Only the week of July 2–8 showed a trend toward earlier break-up in Mittimatalik region with an $R^2 = 0.34$ ($p < 0.5$). There is also a high amount of variability in sea ice break-up, and earlier break-up has become more frequent in the last 10 years (**Figure 12**). The variability in weekly break-up was not as large compared to freeze-up (**Figure 13A**). For the first 3 weeks of break-up, variability is minimal: June 18–24 has an IQR of

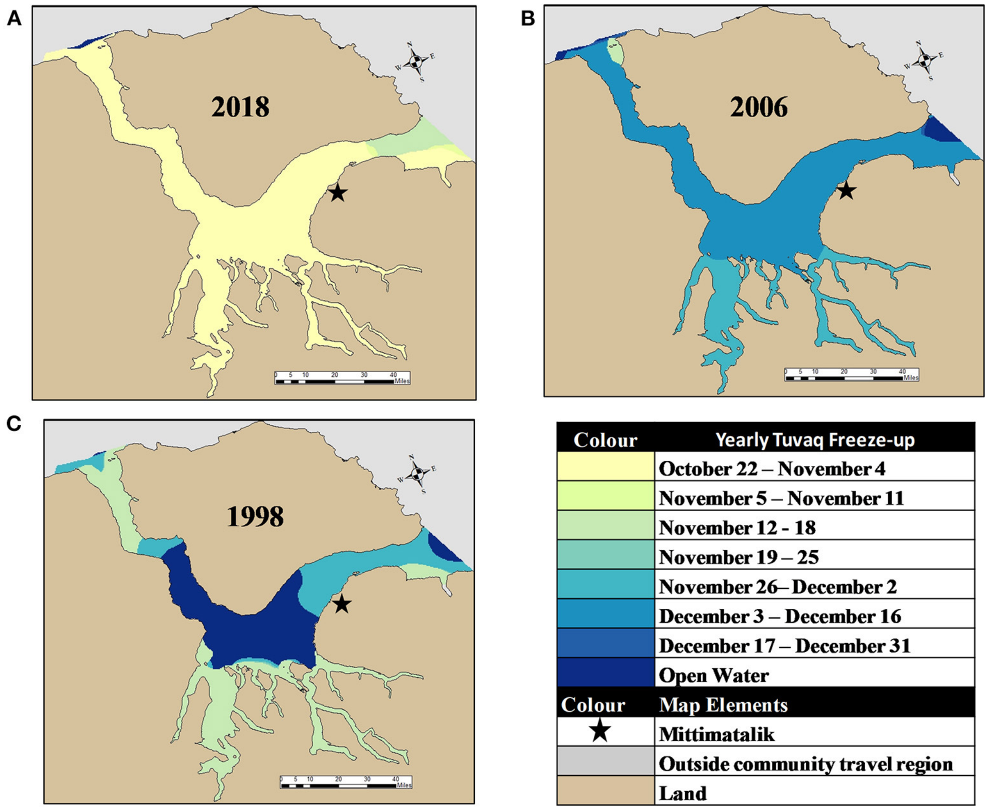

FIGURE 8 | Yearly maps showing the spatial formation of tuvaq for the Mittimatalik region. **(A)** Example from 2018 showing the weekly freeze-up spatial pattern during years when the tuvaq formation was unusually early. **(B)** An example from 2006 showing of the weekly freeze-up spatial pattern during years when the tuvaq formation was unusually late. **(C)** The one exception to the normal freeze-up pattern in 1998 when tuvaq formed last in Eclipse Sound.

3%; June 25 to July 2 an IQR of 10%; and July 2–8 an IQR of 12%. The outliers for the week of July 2–8 correspond to the 2016 and 2019 seasons that broke up unusually early. The 2019 season had the earliest break-up on record with 97% of the region broken-up by July 9–15 (see **Figure 14A** for 2019 map). At the mid-point of break-up, variability increases with the weeks of July 9–15 and July 16–22 having IQRs of 34 and 24%, respectively (**Figure 13A**). The outlier for the week of July 16–22 corresponds to the 2002 season, with only 32% of the sea ice broken-up this week. The final week of break-up, July 23–29, had minimal variability with an IQR of 3%. The outliers for the week of July 23–29 correspond to the years of 1999 and 2005. The year of 2005 had the latest break-up in our record with only 64% of the sea ice broken-up this week (see **Figure 14B** for 2005 map).

The Navy Board sinaa has been breaking up earlier in the last 10 years. For example, 2011, 2013, and 2016 represent the earliest break up years in our 23-year record (**Figure 10C**). The trend for the Navy Board sinaa had an $R^2 = 0.18$ ($p < 0.05$; **Figure 10C**). When compared to the two earliest tuvaq break-up years of 2016 and 2019, the Navy Board sinaa responded in 2016 with the earliest break-up date in our record (July 01). However, for 2019, the Navy Board sinaa break-up date was near normal around July 15th. Sikumiut have also discussed that the Tursukattak sinaa is not as stable as it has been in the past. The Tursukattak sinaa shows a moderate trend for earlier break-up in July with an $R^2 = 0.42$ ($p < 0.05$; **Figure 10B**). The Tursukattak sinaa broke-up early in the anomalous years of 2016 and 2019. In 2016, it broke around July 10 and in 2019 around July 7, the earliest break-up date for this sinaa in the record.

The sinaangit can fracture and sections of tuvaq can break off to form a new sinaa during the break-up season (**Figure 14**). The yearly maps were analyzed to understand if the Tursukattak sinaa fractures and retreats to any consistent locations during break-up. The Tursukattak sinaa fractured to a variety of locations; however, in 17 out of 23 years (74% of the time), it did fracture to a location called Ukkuanguaq (**Figure 14**). Additionally, in 16 out of these 17 years, Ukkuanguaq is the last location of the Tursukattak sinaa before the tuvaq completely breaks-up.

The outlier break-up years from **Figure 13A** were visually analyzed for any differences in spatial patterns for where and when the sea ice broke-up. The patterns were consistent with the seasonal spatial evolution of the average break-up maps in **Figure 11**. However, Arreak explained that in some years, the sea ice in front of the community can break-up earlier than at the Tursukattak sinaa. To continue to hunt and fish as long as possible, Mittimatalingmiut will travel overland to access

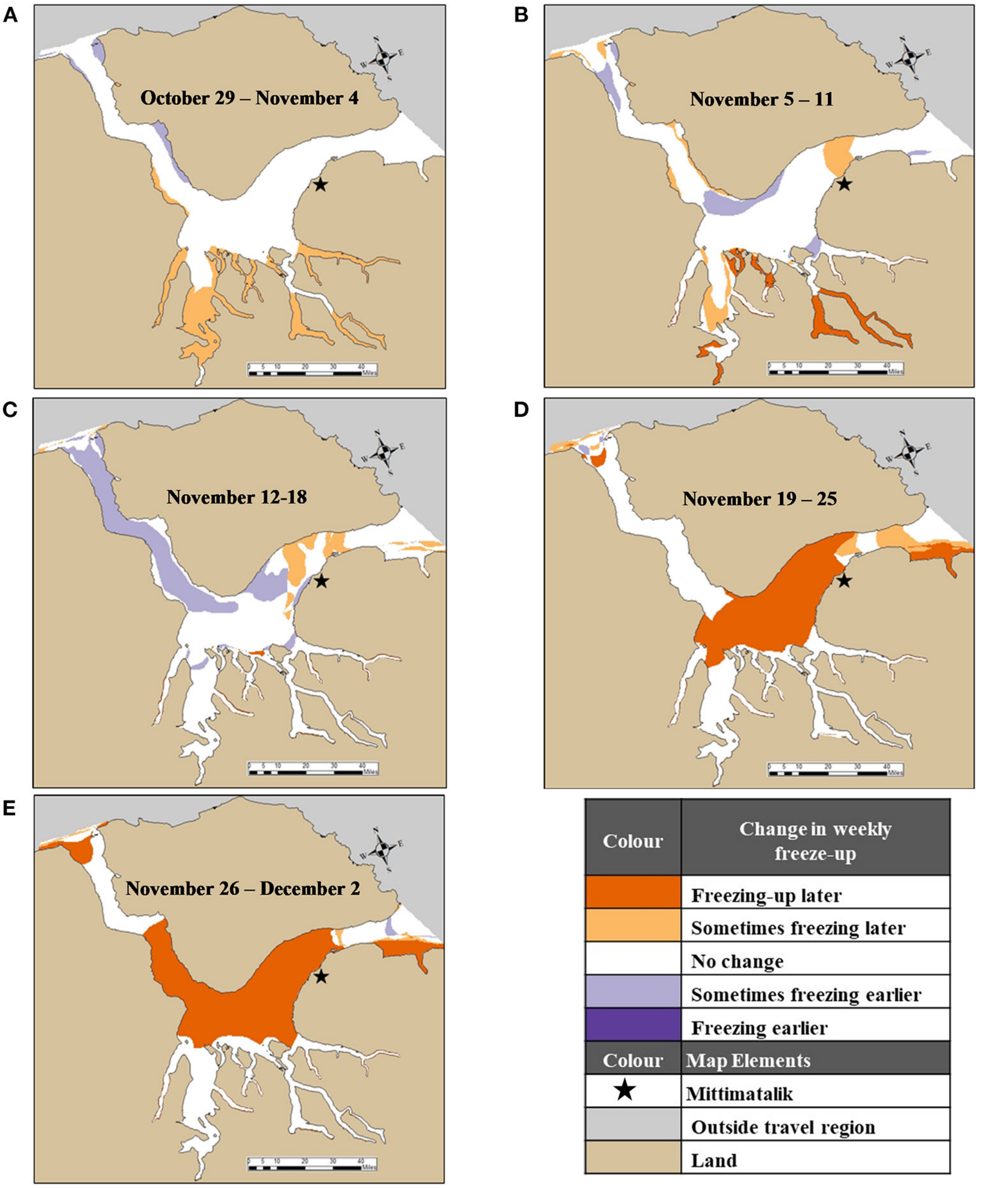

FIGURE 9 | Weekly difference maps showing areas where tuvaq is forming earlier or later in the last 10 years (2010–2019). **(A)** Difference map, October 29–November 4. **(B)** Difference map, November 5–11. **(C)** Difference map, November 12–18. **(D)** Difference map, November 19–25. **(E)** Difference map, November 26–December 3.

the sea ice just past Igarjuaq (Mount Herodier; **Figure 1**). The average break-up maps did not capture this pattern, so we again visually reviewed the individual yearly maps. This type of break-up pattern occurred 11 out of 23 years, just less than half of the time (48%) in the years of 1998, 1999, 2000, 2003, 2006, 2007, 2009, 2011, 2015, 2018, and 2019 (see **Figure 14C** for 2006 map). This pattern of break-up was fairly random and there was no increase in the frequency of this pattern of break-up in the last 10 years. Finally, we examined whether the spatial and temporal patterns of sea ice freeze-up in the fall influences sea ice break-up patterns in late spring, but no obvious patterns were detected.

To understand the critical periods for sea ice break-up, the weeks with the highest percentages of break-up were extracted for each year from 1997 to 2019. **Figure 13B** shows that a majority of break-up is distributed over a 3-week period from July 9 to 29. The week with the highest average percentages of break-up was July 16–22, in which almost half of the annual break-up occurs (48%).

The weekly difference maps show spatially where sea ice break-up is changing the most in the last 10 years of the climatology (2010–2019; **Figure 15**). The June 25-July 1 and July 2–8 difference maps show that the sea ice is breaking up earlier in: the sounds and inlets; at river mouths; in front of Mittimatalik;

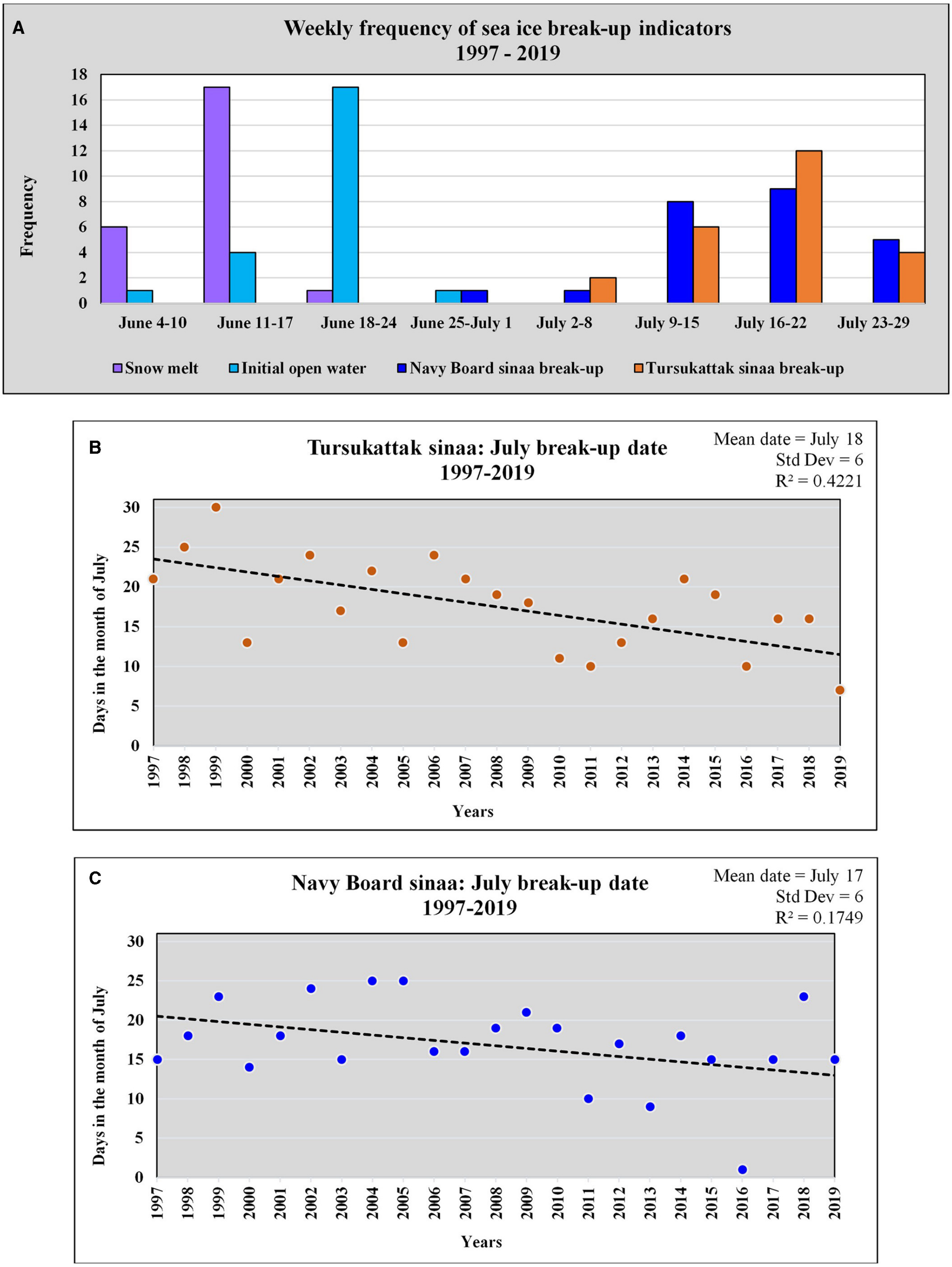

FIGURE 10 | (A) Frequency graph for indicators of break-up, 1997–2019. **(B)** Graph showing the Tursukattak sinaa July break-up dates, 1997–2019. **(C)** Graph showing the Navy Board sinaa July break-up dates, 1997–2019.

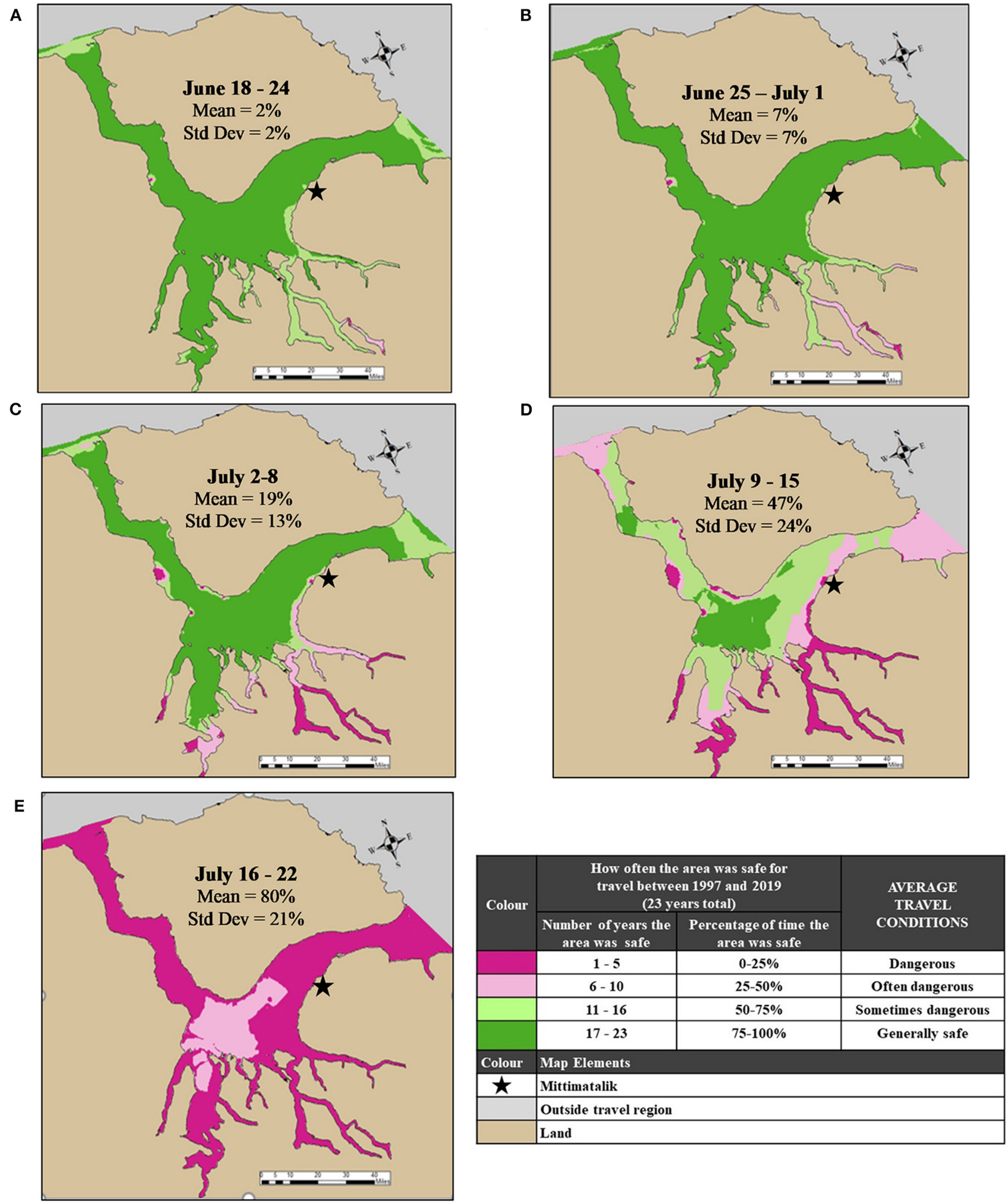

FIGURE 11 | Maps showing weekly average break-up conditions June 18 to July 22, 1997–2019. **(A)** Average break-up, June 18–24. **(B)** Average break-up, June 25–July 1. **(C)** Average break-up, July 2–8. **(D)** Average break-up, July 9–15. **(E)** Average break-up, July 16–22.

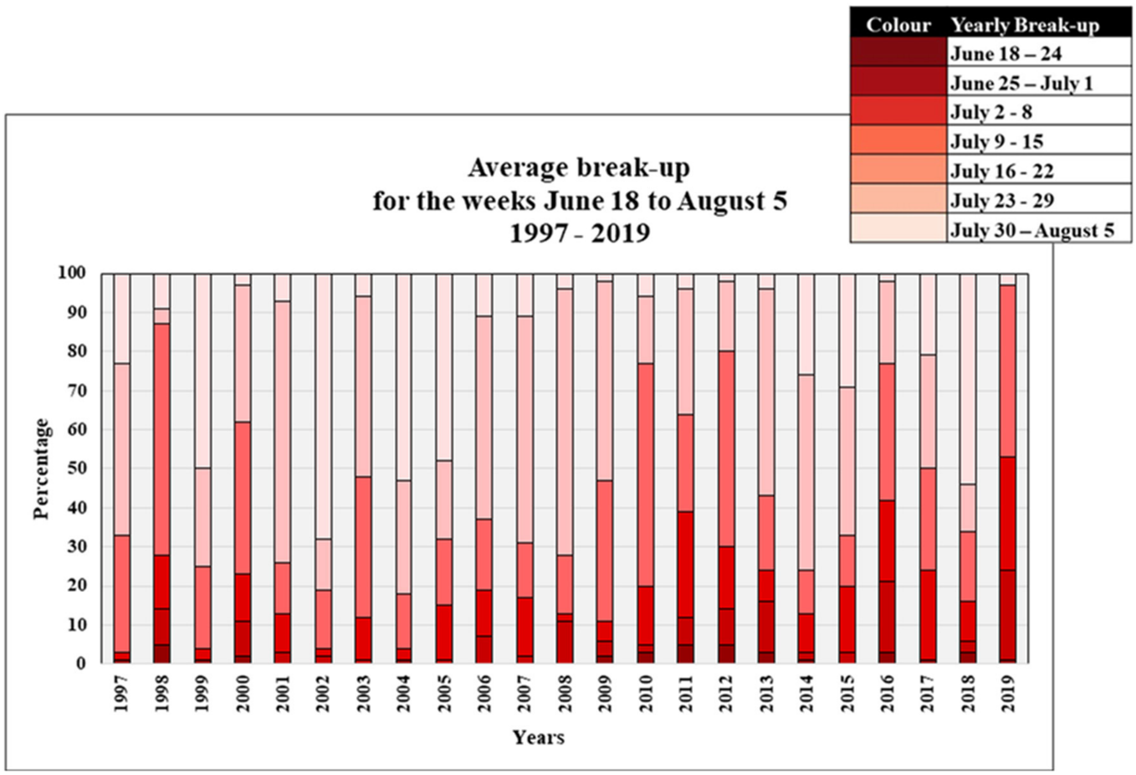

FIGURE 12 | Summary of average break-up between June 18 to Aug 5, 1997–2019. Each bar is a year showing the weekly percentage of break-up by color: dark red for late June; medium red for early July; and light red for the end of July. Years with darker red represent years that broke-up early. Years with more light red represent years that broke-up late.

and at the northern tip of the Tursukattak sinaa (**Figures 15A,B**). The July 9–15 and July 16–22 difference maps show greater break-up in Milne Inlet and Tursukattak (**Figures 15C,D**). The July 16–22 difference map also shows a greater amount of break-up occurring this week in Milne Inlet and Tasiujaq. The July 23–29 difference map shows no spatial changes in sea ice break-up during the last 10 years (**Figure 15E**).

DISCUSSION

The Mittimatalik siku asijjipallianinga not only documents trends, spatial patterns and locations of sea ice change in the Mittimatalik region, but it also addresses community-identified questions from an Inuit point of view, and at spatial and temporal scales that assessments such as the IPCC SROCC currently cannot address. Our discussion first looks at the benefits of this IQ-based based climatology and its application for community and regional sea ice travel safety. We then discuss the value of this IQ-based sea ice climatologies to meet their Mittimatalingmiut environmental assessment needs.

IQ-Based Research for Community Adaptation Needs

It is important to note that this research is not an example of integrating or incorporating IQ into western science. These approaches tend to select IQ that fits or validates western research questions (Bravo, 2009; Bohensky and Maru, 2011; ITK, 2016; McGrath, 2018). In this IQ-based sea ice climatology, we turned typical research approaches inside out by utilizing western science data sources to apply IQ to Inuit research questions (Bell, 2016). In this project, the satellite imagery and CIS charts were used to apply Sikumiut's IQ to the reconstruction of a 23-year ice climatology at seasonal to weekly scales. Additionally, IQ determined the approach to the analysis, filled gaps in the analysis and in the interpretation of the results to answer Mittimatalingmiut sea ice adaptation needs.

Arreak's teachings and travel experience allowed him to interpret the sea ice break-up in the satellite imagery based on his IQ and from an Inuit travel safety perspective. He was able to identify in the satellite imagery early signs of melt and aajurait in the satellite imagery that would have remained undetected without this context specific IQ and on-ice experience. Arreak digitized the locations of hundreds of aajurait over the 23 break-up seasons. In our GIS analysis, we were unable to find any spatial or temporal patterns for where and when, or if specific aajurait were key locations for break-up. However, in the IQ workshops Sikumiut mapped the main locations of the re-occurring aajurait without hesitation (**Figures 2A**, **3A,B**). Additionally, Sikumiut already knew of the significance of the Ukkuanguaq aajuraq, but being able to quantify that the Tursukattak sinaa fractures and retreats to this location 74% of the time supports community sea ice adaptation needs. For example, talks are already underway

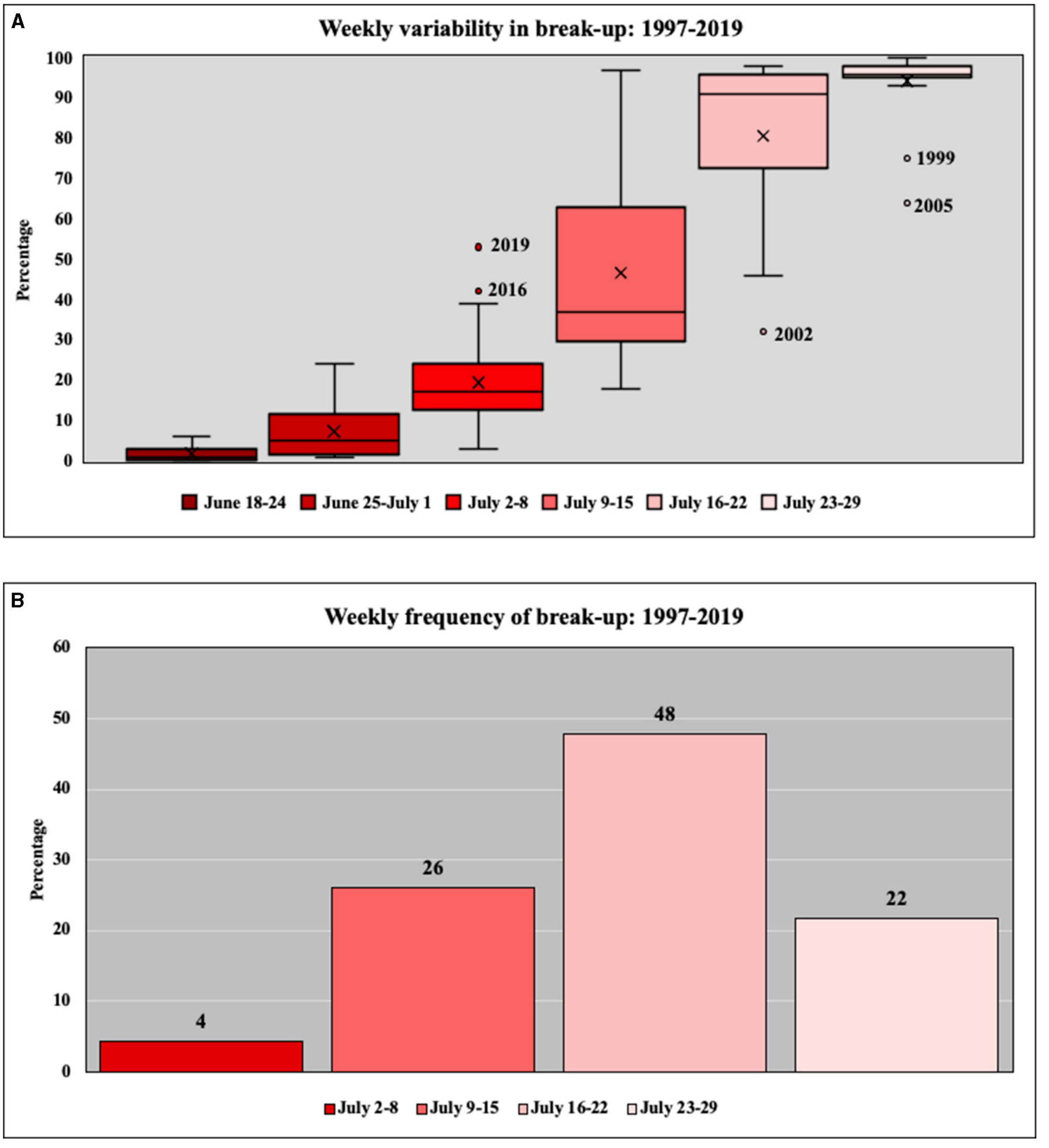

FIGURE 13 | (A) A summary of the weekly variability in break-up from 1997 to 2019. The box outlines the interquartile range, the average range in the variability of break-up for each week over the 23-year period (1997–2019). The line through the box is the median and the X denotes the mean. The vertical "whisker" lines show the minimum and maximum values. The dots correspond to outliers, or years with unusual break-up percentages. **(B)** Weekly frequency of break-up, 1997–2019.

to position time-lapse cameras and other monitoring equipment at this location to provide Mittimatalingmiut advance notice of break-up (Bell et al., 2020).

Arreak also pointed out that the average and difference break-up maps did not capture the years when the sea ice in front of the community breaks-up earlier than at the Tursukattak sinaa. This is an important break-up pattern that occurred 11 out of 23 years, 48% of the time (**Figure 14C**). Without Arreak's IQ, this break-up pattern would have been missed if we relied solely on statistical and GIS analyses. When you factor in that the sea ice is breaking up earlier (**Figure 15**) with the fact the sea ice in front of the community breaks-up first 48% of the time, access to the Tursukattak sinaa is becoming extremely difficult in late June and early July. Within the community, there have been suggestions to build a road to Igarjuaq as an adaptation strategy to maintain consistent access to the Tursukattak sinaa (**Figure 1**).

Sikumiut validated the average weekly break-up maps to ensure that the maps aligned with their IQ (**Figure 2B**). The

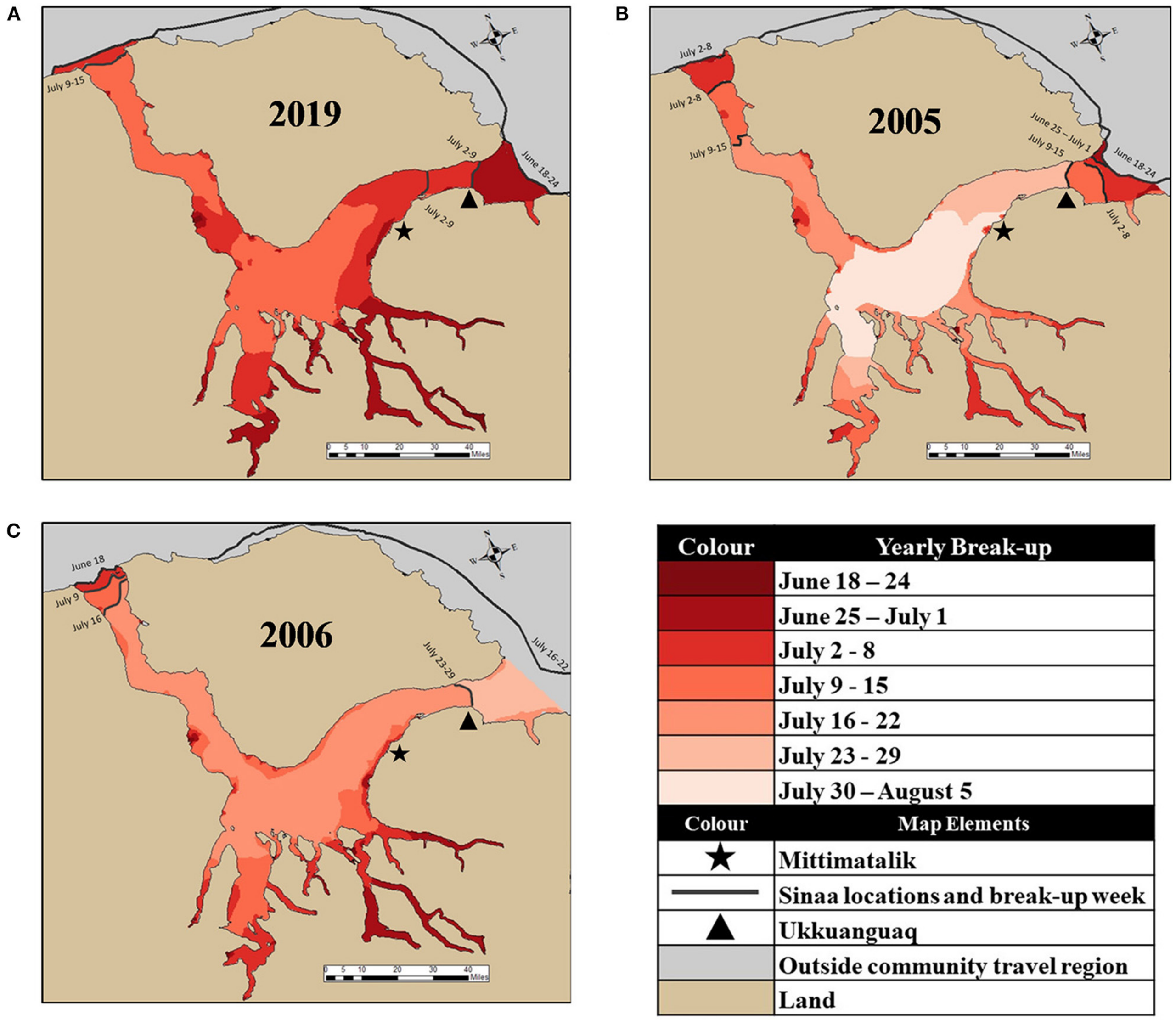

FIGURE 14 | Yearly maps showing the spatial break-up of sea ice for the Mittimatalik region. **(A)** Example from 2019 showing the weekly spatial pattern for an unusually early break-up. **(B)** Example from 2005 showing the weekly spatial pattern for an unusually late break-up. **(C)** Example from 2006 showing the weekly spatial pattern when the sea ice at the Tursukattak sinaa breaks last.

benefit of the weekly average and difference maps are that they document and mobilize Sikumiut's knowledge from a seasonal to a weekly scale and highlight areas that have become more dangerous for sea ice travel. During break-up, these weekly maps can support travel planning. For example, by the week of June 25 to July 1, Mittimatalingmiut need to be cautious when traveling in Tay Sound because on average, the sea ice is sometimes dangerous (**Figure 11B**). By the week of July 2–8 travel in Tay Sound is frequently dangerous (**Figure 11C**), but based on the increase in break-up in the last 10 years, this area should sometimes breaks-up early and should be avoided (**Figure 15**).

When you view the Sikumiut seasonal sea ice IQ spring travel map (**Figure 3B**) compared with the weekly average break-up maps (**Figure 11**), you will notice striking similarities in the dangerous travel areas. However, the Sikumiut map shows additional hazardous sea ice areas along the southeast shore of *Sirmilik* (Bylot Island; **Figure 1**), around the Tursukattak sinaa, and the main aajurait locations not captured in the weekly average maps. To fill these gaps, the final version of the weekly average maps will overlay Sikumiut's additional IQ of aajurait and hazardous travel area locations.

The Sikumiut seasonal sea ice IQ winter travel map shows travel conditions once the sea ice has become tuvaq, in other words when it is generally safe for travel (**Figure 3A**). Early winter sea ice travel requires extreme caution and Sikumiut recommends that only the most knowledgeable and experienced hunters break initial snowmobile trails. Sikumiut would not historically have the bird's eye perspective of the region provided

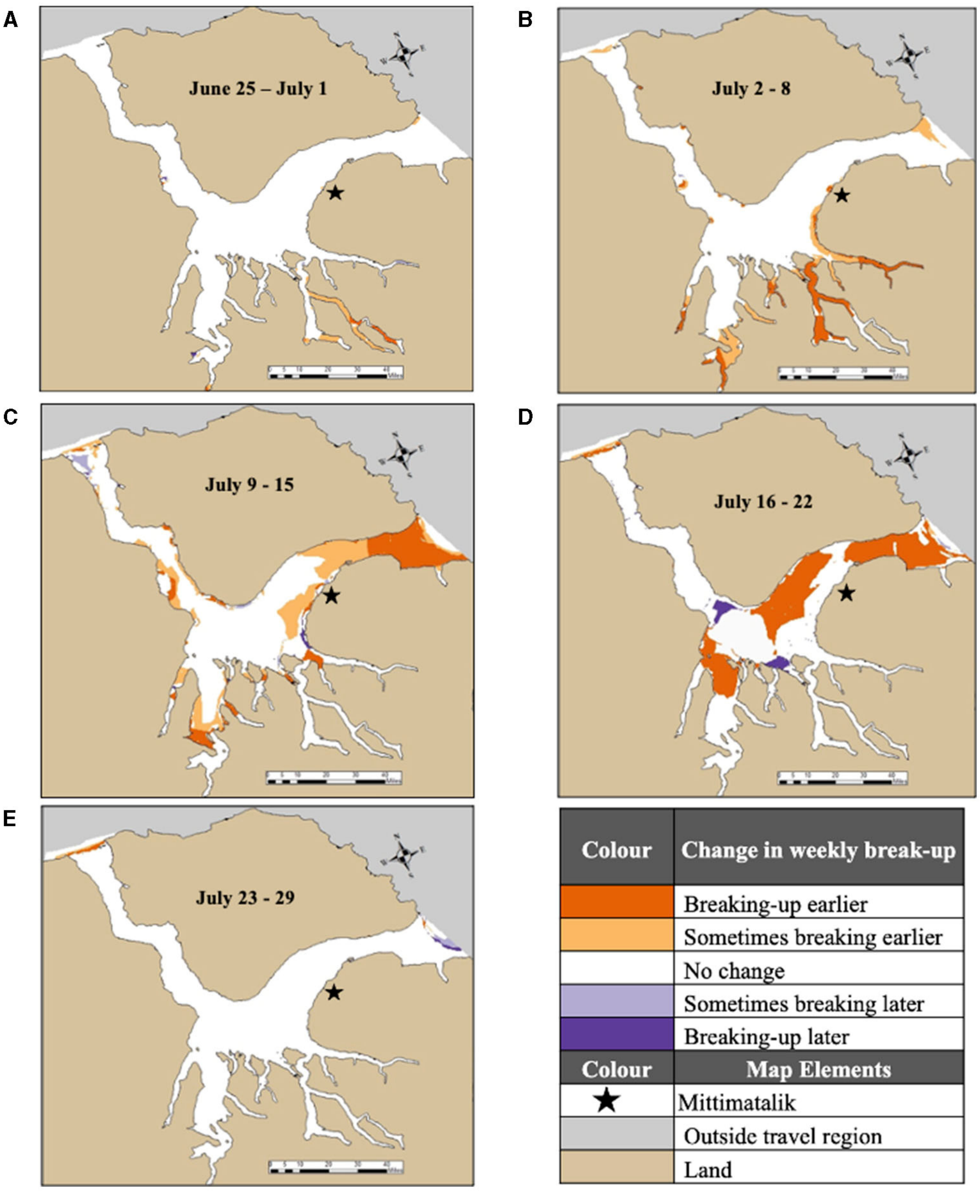

FIGURE 15 | Weekly difference maps showing areas where break-up is occurring earlier or later in the last 10 years (2010–2019). **(A)** Difference map, June 25–July 1. **(B)** Difference map, July 2–8. **(C)** Difference map, July 9–15. **(D)** Difference map, July 16–22. **(E)** Difference map, July 23–29.

by the satellite data to monitor tuvaq formation. Sikumiut's freeze-up IQ is based on experiences passed down through generations on where it is normally safe to access the sea ice from the land in early winter. For example, Sikumiut members know that on average the first areas of tuvaq formation are in the southern inlets and sounds (**Figure 5B**). However, the weekly average tuvaq maps for freeze-up, based on the CIS ice charts, show the formation of tuvaq in Navy Board inlet, normally inaccessible for Mittimatalingmiut until the ice is safe for travel in Tasiujaq. Sikumiut reviewed and validated these maps to support travel planning in late freeze-up. For example, the November 12–18 map shows that the sea ice is normally not safe for travel anywhere near the community this week (**Figure 5C**). By the week of November 26 to December 2 it is normally safe to travel on the sea ice from the community into Tasiujaq, but it is normally still not safe for sea ice travel in Tursukattak until the end of December (**Figure 5E**). Once more the Sikumiut winter seasonal map provides additional detail, such as naggutiit, ivujuk, and siku saattuq aragulimaamik not in the weekly freeze-up maps. To fill these gaps, the final versions of the weekly maps will overlay the locations of these Sikumiut features to enhance the sea ice travel safety information for freeze-up.

In Canada's north, search and rescue operations are a complement of multi-jurisdictional partners. In Nunavut communities, local volunteers in Mittimatalik are often the first responders. Nunavut Emergency Management (NEM) coordinates at the Territorial scale. Based on the severity and type of the incident, NEM can request support from the following Federal agencies: Department of National Defense (air); Royal Canadian Mounted Police (land); and Canadian Coast Guard (sea). The Mittimatalik siku asijjipallianinga can also support the safety and situational awareness of regional and national search and rescue partners that would have a limited knowledge of the area and local sea ice conditions. For example, hazardous sea ice areas and areas of shelter to focus search and rescue efforts. The weekly average maps would support the effective, efficient and proactive deployment of resources and assets (human or infrastructure based) based on known areas of high risk at a weekly scale. Additionally, community scale IQ-based sea ice climate maps would be beneficial for national ice services. The presence of melt ponds in the spring saturates the SAR imagery making it impossible to identify sea ice features. As well, spring storms with significant cloud cover can result in weeks without optical imagery. Ice services would benefit from such community scale climate atlas' to help fill in satellite imagery gaps during the sea ice break-up season.

IQ-Based Research for Environmental Assessments

The normal open water season for shipping to the Mary River mine is from August 5 to October 15 (Bourbonnais et al., 2016). In 2020, BIM requested an extension to the shipping season from approximately July 15 to November 15, based on declining sea ice extent in the Arctic. An ice conditions shipping assessment report was submitted to the Nunavut Impact Review Board (NIRB) describing current shipping conditions to and from the mine (Bourbonnais et al., 2016). The ice conditions report highlights that climate change is resulting in sea ice freezing up later and breaking up earlier in the Canadian Arctic (Bourbonnais et al., 2016). The ice conditions report also outlined that the sea ice conditions in the region are highly variable, that climate change increases the risk of dangerous mobile old ice floes, and that ice-breaking support would be needed to ship during these shoulder seasons (Bourbonnais et al., 2016).

Responses to the proposed BIM lengthening of the shipping season have been sent from Sikumiut, the Mittimatalik Hunters and Trappers Organization (MTHO) and the Canadian Department of Fisheries and Oceans (DFO) to NIRB. All outline the importance of sea ice in the fall and late spring for wildlife reproduction and migration, and concerns regarding the impacts of noise from icebreaking on marine mammals (DFO, 2019; MTHO, 2021; Sikumiut, 2021). Sikumiut and the MTHO both outline the importance of sea ice for their culture and food security. They also emphasize that their concerns are based on IQ and that the environmental assessment process has not given IQ an equivalent voice when understanding the impacts of an extended shipping season on Mittimatalingmiut (MTHO, 2021; Sikumiut, 2021). Although NIRB outlines that their process is guided by IQ principles and that IQ has an important contribution to make to the review process (NIRB, 2021), it has been very difficult for oral knowledge to compete with technical reports and in evidence based decision-making processes (White, 2006; Healey and Tagak, 2014; McGrath, 2018).

The Mittimatalik siku asijjipallianinga provides IQ-based evidence concerning the proposed extended shipping seasons, and raises some interesting questions. For example by the week of November 12–18, Milne inlet averages 75–100% tuvaq and by November 19–25, there is 50–75% tuvaq in northern Tasiujaq, which would require a considerable amount of icebreaking to ship through (**Figures 5B,C**). **Figure 7A** also shows that a majority of tuvaq formation (56%) occurs in the first 2 weeks of November. Shipping during this critical period could compromise the formation of tuvaq and the Tursukattak sinaa, consequently affecting winter sea ice travel and wildlife. It is interesting to note that both the ice conditions report (Bourbonnais et al., 2016) and the siku asijjipallianinga used the CIS charts to review freeze-up conditions. However, the shipping report interpreted the data from a safe shipping perspective and the siku asijjipallianinga from a safe sea ice travel and wildlife perspective. While the shipping report notes that there is an expectation that the sea ice extent in Mittimatalik is declining due to climate change, we found no trend toward later freeze-up, and that in the last 10 years tuvaq freeze-up could be occurring earlier in some areas. The Milne Inlet port shows signs of earlier tuvaq freeze-up during the week of November 5–11 (**Figure 5B**), which could have implications for the feasibility of extended shipping at the port. Due to the high variability of freeze-up conditions (**Figure 7A**), it is impossible to pre-determine a specific week to cease shipping for the season. Sikumiut have recommended that the end of the shipping season be assessed on a year-by-year basis, according to the sea ice conditions at the time (Sikumiut, 2021).

The Mittimatalik siku asijjipallianinga also evaluated the potential impacts to sea ice travel based on the proposal to start shipping earlier around July 15. On average, by the week of July 16–22, the Mittimatalik region is 80% broken up (**Figure 11E**) and normally the Navy Board and Tursukattak sinaangit break-up this week (**Figures 10B,C**). Also for this week there is a trend toward an earlier break-up of the Tursukattak sinaa ($R^2 = 0.42$), and along the shipping route to Milne Inlet in the last 10 years (**Figure 15D**). However, the break-up conditions are variable (**Figure 13A**). For example, even in the two most recent years in the record, Mittimatalingmiut experienced both an early (2019, 97% break-up by July 9–15) and late (2018, 95% break-up by July 23–29) break-up (**Figure 11**). Shipping earlier into the first 2 weeks of July would compromise community sea ice access to the Tursukattak sinaa in years when they are experiencing a late break-up. A follow-up letter from Sikumiut to NIRB is being sent to highlight this IQ-based evidence from the Mittimatalik siku asijjipallianinga in preparation for the next round of hearings.

CONCLUSION

Mittimatalik is just one out of 48 coastal communities in Inuit Nunangat that need answers to their climate change questions.

International assessments such as the IPCC SROCC cannot address community-scale issues based on the current global scale of the models and methodologies used. The community of Mittimatalik is already dealing with the impacts of climate change influencing sea ice conditions, compounded by the pressure to increase shipping into the margins of the sea ice travel season. A deep climatological history of sea ice continues to thrive in IQ, but for many Inuit communities, it has yet to be documented. In the Mittimatalik siku asijjipallianinga, IQ was the foundation upon which their sea ice climatology was built. While satellite imagery, CIS ice charts and other western methods were used to document and mobilize this knowledge from a seasonal to weekly time scale, IQ was the ultimate scientific authority in this project. This ensured that the data were analyzed from an Inuit travel safety perspective, and according to an intimate knowledge of the local environmental conditions. As a result, this IQ-based research was able identify greater detail in the supporting data, fill gaps in the data, and provide direction on how interpret the data to reveal patterns that western-based research methods could not capture.

This atlas provides an adaptation tool that Mittimatalingmiut can use for safe sea ice travel planning, for monitoring specific sea ice indicators during break-up, and in planning alternative land routes in late spring to maintain access to the Tursukattak sinaa. These maps can also support the safety and situational awareness at regional scales for search and rescue partners that would have limited knowledge of local sea ice conditions. This project provides a practical example for how to develop an IQ-based sea ice climatology, and how this research approach can serve local Inuit community needs and beyond at regional scales. There would be a great benefit in expanding this work to other Inuit communities to support local safe sea ice travel and emergency management programs and practices across the Canada North. This atlas also has great value to the larger scientific community as climate change does not affect all areas of the Arctic equally.

The Mittimatalik siku asijjipallianinga demonstrates the scientific merit of IQ and its value in environmental assessments. The IQ-based evidence from the atlas shows that extending the shipping season into the first 2 weeks of November and the first 2 weeks of July will compromise the integrity of the sea ice for safe travel, and wildlife migration and reproduction. If shipping is extended into the freeze-up and break-up seasons to support mining activities, Mittimatalingmiut now have a baseline of their local sea ice conditions with which to compare and provide evidence for any future cumulative effects.

This co-produced research is also an example of the time required to meaningfully engage and work with Indigenous knowledge holders, whether its for environmental or scientific assessments like the IPCC SROCC. It required an investment of over 4 years in which Inuit were involved in the discussions from the very beginning and throughout the research, not just during a couple of workshops. By co-producing the research together and agreeing from the beginning on how to collect, analyse, and interpret the information, different knowledge systems can work together to address community-scale issues missing in IPCC SROCC reports.

DATA AVAILABILITY STATEMENT

The datasets presented in this article are based on Indigenous knowledge and will require permission from the Sikumiut Committee. Requests to access the datasets should be directed to Katherine Wilson, katherine.wilson@mun.ca.

ETHICS STATEMENT

Written informed consent was obtained from the Sikumiut Committee for the publication of potentially identifiable images and data included in this article in the Memorial University-Sikumiut research agreement (Wilson, 2018).

AUTHOR CONTRIBUTIONS

The Sikumiut Committee, AA, and KW contributed to conception and design of the study. KW archived the data, organized the database, performed the statistical analysis and map layout, and wrote all drafts of the manuscript. AA performed the satellite interpretation, digitization, and GIS spatial analysis. All authors contributed to the methodological design, as well as manuscript review and revisions.

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The Ocean and Cryosphere in a Changing Climate in Latin America: Knowledge Gaps and the Urgency to Translate Science Into Action

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Climate Change hazards to social-ecological systems are well-documented and the time to act is now. The IPCC-SROCC used the best available scientific knowledge to identify paths for effective adaptation and mitigation of climate change impacts on the ocean and cryosphere. Despite all the evidence highlighted by SROCC and the key role of the ocean and cryosphere for climate change at all levels, Latin America (LA) faces challenges to take effective action mostly due to socio-economic vulnerability, political instability and overall technical capacities. Countries have adopted diverse actions as the information needed by policy makers has been made available, not necessarily in accessible and inclusive ways. Regional imbalance in economic development, technological level, capacity development, societal involvement, and governmental oversight have contributed to skewed geographical and technological gaps of knowledge on key ecosystems and specific areas preventing effective climate actions/solutions. We analyze the Nationally Determined Contributions (NDCs) from the region as proxies to the incorporation of IPCC recommendations. The gaps and opportunities for the uptake of ocean and climate science to political decision making is discussed as five key aspects: (i) climate assessment information and regional policies, (ii) knowledge production, (iii) knowledge accessibility, (iv) knowledge impact to policy, and (v) long term monitoring for decision making. We advocate that the uptake of SROCC findings in LA policies can be enhanced by: (a) embracing local realities and incorporating local, traditional and indigenous knowledge; (b) empowering locals to convey local knowledge to global assessments and adapt findings to local realities; (c) enhancing regional research capabilities; and (d) securing long-term sustainable ocean observations. Local and regional participation in knowledge production and provision enhances communication pathways, climate literacy and engagement which are key for effective action to be reflected in governance. Currently, the lack of accessible and inclusive information at the local level hampers the overall understanding, integration and engagement of the society to mitigate climate effects, perpetuates regional heterogeneity and threatens the efforts to reverse the course of climate change in LA. Local researchers should be empowered, encouraged, rewarded and better included in global climate-ocean scientific assessments.

Keywords: climate change, SROCC, local knowledge, policy makers, Latin America

INTRODUCTION

The critical importance of the ocean and the cryosphere to the climate system (Reid et al., 2009), hydrological cycles (Schanze et al., 2010; Liu et al., 2020) and the consequences to society (Nicholls, 2010) stimulated the IPCC to commission a Special Report on the Ocean and Cryosphere in a Changing Climate (hereafter SROCC) (IPCC, 2019a), which assessed climate change impacts on marine and coastal ecosystems. The combined effects of ocean warming, ocean acidification, and deoxygenation are reducing primary production and marine biodiversity, and impacting associated ecosystem services such as nutrient cycling, carbon sequestration and fisheries (Bindoff et al., 2019). Sea level rise and ocean extreme events are causing coastal damage, with natural and capital losses, and many associated socio-economic impacts (Nicholls, 1995; Leatherman, 2001). People dependent on or living in close connection with coastal, polar and mountain environments are especially vulnerable to the hazards of ocean and cryosphere change (Oppenheimer et al., 2019).

Climate change is a global phenomenon, but the scale to act is local, primarily influenced by a country's policies, geography, socio economic development, and vulnerability to climate-risks. Consequently, climate coastal adaptation policies have been developed, with substantial variations between countries, and across developmental status (Klein et al., 1999, 2001). Investigating how scientific knowledge and international recommendations based on science are being taken at national level is not simple, due to the time-lag between implementation, monitoring, evaluation and reporting. In addition, climate change adaptation actions generally lack scientific uptake and on-the-ground change, with most focus being on assessing vulnerability, compared to developing plans and actions (Gibbs, 2015).

Latin America and the Caribbean[1] (hereafter LAC) represents a high contrast region where wealth and prosperity coexist with vulnerability and extreme poverty, explained by low growth (Fernández-Arias and Fernández-Arias, 2021). The region hosts 1/3 of the world's most biodiverse countries and highly urbanized regions (UNEP, 2011). It comprises 46 countries, dependent territories and overseas departments on the edge of the Atlantic and Pacific oceans and the Caribbean Sea and is limited to the south by the Southern Ocean. Altogether, LAC has more than 30,000 km of coastline, ranging from the tropical region—dominated by mangroves, seagrass meadows and coral reefs—to subtropical and temperate areas, dominated by salt marshes, rocky shores and macroalgal beds all the way to the Drake passage, where the influence of the Southern Ocean and Antarctica is well-documented (Sijp and England, 2004; Scher and Martin, 2006; Livermore et al., 2007; Yang et al., 2014; Viebahn et al., 2016; England et al., 2017). This continent-wide latitudinal range reflects diverse oceanic domains and climate influences, responsible for the high diversity of marine habitats and ecosystems (Miloslavich et al., 2011; Turra et al., 2013; Spalding et al., 2017) where hotspots of "exceptional marine biodiversity" and fisheries coincide with areas most severely affected by global warming (Ramírez et al., 2017). Coastal areas surrounding the Mesoamerican reef and nearby islands are low-lying and extremely vulnerable to sea level rise. Extreme hydrometeorological events are frequent and coastal erosion is a widespread threat (particularly severe in Northern and Northeast Brazil—Silva et al., 2014), associated with human interventions, poor coastal planning and management but also influenced by the morphodynamic nature of the coast (Silva et al., 2014), as most coastal areas in the region. The diversity of this region is also reflected by a wealth of peoples, languages, social-political systems, cultures, traditions and origins that constitute a unique mosaic of diversities with 780 indigenous peoples and 560 different languages (Freire et al., 2015) that heightens regional imbalance and skewed geographical and technological gaps.

Despite the alarming magnitude and extent of climate change effects shown by SROCC, Latin America (hereafter LA) is challenged to take effective action due to socio-economic vulnerability. Extreme poverty in LA reached unprecedented levels in 2020 (ECLAC, 2021), and social inequality indices, like unemployment and labor participation rates, have worsened particularly among women, despite the recent emergency social protection measures adopted in the COVID-19 pandemic. The recently published Regional Human Development Report[2] for LAC (UNDP, 2021) highlights that concentration of power, violence in all its forms and failed social protection of policies and frameworks cause the contrasts found in the region. These promote high inequality and low growth, challenging the intake of the recommendations from SROCC even though the region plays a key part in the global green recovery (UNDP, 2021).

This paper analyzes the adopted Nationally Determined Contributions (NDCs) in LAC as a reflection of the incorporation

[1] The acronyms LAC and LA are not used interchangeably: LAC refers solely for evidence from Latin America and Caribbean while LA is used when evidence refers to Latin America.

[2] Regional Human Development Report for Latin America and the Caribbean in https://www.latinamerica.undp.org/content/rblac/en/home/library/human_development/regional-human-development-report-2021.html.

of IPCC recommendations. We briefly discuss gaps and opportunities in the region for the uptake of ocean and climate science to political decision making, organized in five key categories: (i) climate assessment information and regional policies, (ii) knowledge production, (iii) knowledge accessibility, (iv) knowledge impact to policy, and (v) long term monitoring for decision-making. We finally present some conclusions and propose future actions.

NATIONALLY DETERMINED CONTRIBUTIONS (NDCS) AS SCIENCE UPTAKE INDICATORS

Nationally Determined Contributions (NDCs) are official Government commitments to comply with UNFCCC's targets. NDCs may also reflect how IPCC findings are perceived and incorporated into policy documents that go beyond current national climate plans and bring us closer to the Paris Agreement goals of decarbonizing economies and improving resilience. We reviewed the NDC reports submitted by 31 Latin American and Caribbean countries to the NDC registry[3] and searched for expressions such as "oceans and coasts," "fisheries," "risk management," "gender," "UN 2030 Agenda," "interculturality," "community-based solutions," "ecosystem-based adaptations," and "cryosphere." In addition, we incorporated Socio Economic indicators such as Gross Domestic Product (GDP) per capita, Human Development Index (HDI), and specific GDPs for 2017 and 2019, relative to the overall values of GDP estimates for the whole world, extracted from the World Bank Database[4] Results are shown in **Table 1**.

Risk management was the most common feature, addressed by 29 countries, followed by Ocean and coastal activities ($n = 27$). Fisheries actions were reported for 19 countries, while ecosystem-based adaptation appeared in 18, such as Mexico's Blue Carbon action. Gender equity/balance issues were a concern for 17 countries and frequently mentioned by most countries, although not included by Guatemala and Uruguay. Community based adaptation is of particular interest in 16 countries, while Agenda 2030 has been considered by 15 countries in their NDCs. Interculturality, the existence and equitable interaction of diverse cultures and the possibility of generating shared cultural expressions through dialogue and mutual respect[5] was important for 13 countries. The least frequent concern was with the Cryosphere, only included in the NDCs of Argentina, Chile and Peru (**Figure S1**).

NDCs in the LAC region have been developed according to local context and capacities. Commitments to climate change mitigation and adaptation are frequent, but specific targets to "oceans and cryosphere" have not been prioritized. For example, SROCC reports coral reefs as amongst the most susceptible ecosystems and yet the Mesoamerican Arrecifal Barrier has not been included among local NDCs. The same rationale goes for "Ocean and Coasts" in the NDCs from Brazil, Guyana, Nicaragua, and Peru which seems invisible to national commitments despite the proportion of coastal and marine areas in these countries. NDCs in the region have not yet been impacted by SROCC findings as reflected by the first (and now the second more recent) round of NDC submissions. Unless the climate and ocean communities recognize LAC's socio-economic contexts and associated environmental and social vulnerabilities to consider uniting to act, this scenario might not change significantly over time.

GAPS AND OPPORTUNITIES IN LATIN AMERICA

Climate Assessment Information and Regional Policies

National Adaptation Plans (NAPs) are policy driven commitments that translate the NDCs into local and sectoral actions. Technology Needs Assessments (TNAs) are rights of States to claim the necessary technology to comply with IPCC recommendations. NDCs, NAPs and TNAs are three different, but complementary, instruments that countries in LAC seek to implement. While NDCs are designed to fulfill international commitments, NAPs and TNAs reflect national capacities and local vulnerabilities, yet to be targeted in IPCC assessments.

IPCC assessments are geographically and disciplinarily skewed, strongly based on the most influential science produced by developed countries (Vasileiadou et al., 2011), with a disproportionate influence of formally educated and economically advantaged groups (Castree et al., 2014). Thus, as LAC contributions to SROCC have been limited (nine authors from seven countries), the resulting recommendations also lead to limited local application. Moreover, political leadership in the region favors socio-economic policies over environmental protection (e.g., Custer et al., 2018, in relation to the UN 2030 Agenda). Yet NAPs generally show two main pathways: while developed countries focus on economic risks and opportunities, developing countries prioritize natural resources and conservation (Alves et al., 2020). A clearer connection between environmental threats and socio-economic concerns must be established so regional leaders feel safer and supported to make decisions. Local researchers should be encouraged to work closely with communities and aid in bridging knowledge gaps.

The IPCC epistemic community defines knowledge as information published in peer-reviewed papers, generally neglecting publications in other languages and other sources of knowledge (traditional, indigenous, and local knowledge—ILK). SROCC has made an effort to include ILK (Abram et al., 2019) and yet LA-ILK's representation was slim. One can argue that SROCC has favored traditional, formally educated, and economically advantaged groups as most scientific assessments do (Castree et al., 2014) perpetuating cultural and geographical imbalance. Representation matters and the participation of Lead Authors from LAC in SROCC was also low. Ten out of 103 SROCC authors were self-identified as belonging to 8 LAC

[3]https://www4.unfccc.int/sites/NDCStaging/Pages/All.aspx
[4]https://data.worldbank.org/indicator
[5]As expressed by the Convention on the Protection and Promotion of the Diversity of Cultural Expressions (Art. 4.8).

TABLE 1 | Specific categories listed as targets to meet either UNFCCC's climate targets or specific SDGs from 31 countries from Latin American and the Caribbean region (LAC) according to the NDC Registry*. Countries are alphabetically listed. Economic indicators, *i.e.*, Gross Domestic Product (GDP) per capita, Human Development Index (HDI), and specific GDPs for 2017 and 2019 relative to overall world GDP estimates for these specific years were world extracted from World Development Indicators**, at the World Bank website.

Categories Countries	Risk management	Gender	Ocean and coasts	Agenda 2030	Ecosystem based adaptation	Fisheries	Interculturality	Community based adaptation	Cryosphere	GDP 2017	HDI 2017	R_GDP 2017	R_GDP 2019
Antigua & Barbuda	1	1	1	1						2.17	0.78	1.02	3.00
Argentina	1	1	1	1	1	1	1	1	1	1.76	0.83	0.82	−2.44
Bahamas	1	1	1	1		1				2.08	0.81	0.98	0.17
Barbados	1	1	1		1					0.32	0.80	0.15	−0.18
Belize	1	1	1	1						−0.13		−0.06	−1.27
Brazil	1			1	1		1			0.51	0.76	0.24	0.30
Chile	1	1	1	1	1		1	1	1	−0.24	0.84	−0.11	−0.11
Colombia	1	1		1	1	1	1	1		−0.16	0.75	−0.08	1.49
Costa Rica	1	1	1	1	1	1	1	1		2.80	0.79	1.31	0.89
Cuba	1	1		1						1.77	0.78	0.83	0.00
Dominica	1	1	1			1				−7.00	0.72	−3.28	2.62
Dominican Republic	1	1	1	1	1					3.52	0.74	1.65	3.16
Ecuador	1	1	1	1	1	1	1	1		0.57	0.75	0.27	−1.29
El Salvador	1	1				1				1.74	0.67	0.82	1.48
Grenada	1	1	1		1			1		3.87	0.77	1.81	1.13
Guatemala	1	1				1	1	1		1.36	0.65	0.64	1.78
Guyana	1					1				3.22	0.65	1.51	3.86
Haiti	1	1	1	1						0.96	0.50	0.45	−2.31
Honduras	1	1			1					3.08	0.62	1.45	0.78
Jamaica	1	1	1	1	1	1	1			0.50	0.73	0.23	0.19
Mexico	1	1	1	1	1	1	1	1		0.93	0.77	0.44	−0.91
Nicaragua	1			1	1	1	1	1		3.31	0.66	1.55	−4.03
Panamá	1	1	1	1	1	1	1			3.81	0.79	1.79	1.05
Perú	1		1		1		1	1	1	0.83	0.75	0.39	0.41
Saint Kitts and Nevis		1	1							−2.74	0.78	−1.29	1.65
Saint Lucia	1	1	1	1	1		1	1		2.96	0.75	1.39	0.98
St Vincent and the Grenadines	1	1	1			1				0.66	0.72	0.31	0.12
Suriname	1	1			1	1	1	1		0.76	0.72	0.36	−0.53
Trinidad & Tobago		1								−2.77	0.78	−1.30	−0.29
Venezuela	1	1				1		1			0.76	0.00	0.00

Note that negative scores are shown in red.
https://www4.unfccc.int/sites/NDCStaging/Pages/All.aspx.
***https://ourworldindata.org/human-development-index and http://wdi.worldbank.org/.*

countries[6] (97% of the total authoring contribution). The USA alone had 15 authors, in contrast with other LAC countries such as Cuba and Trinidad and Tobago—the only SIDS regionally represented—as well as Mexico and Brazil—two of the largest and most populated countries in LA—which accounted for one participant each. Apart from language and representation, technical and scientific capacity deficiencies refrained researchers in the region from contributing more (Polejack and Coelho, 2021).

Knowledge Production

Knowledge production of the ocean-climate nexus in LA insofar as the uptake of such knowledge to national advisory exercises is still incipient, possibly due to limited technical capacities in the region combined with a deficient access to marine

[6]These countries are: Argentina, Brazil, Costa Rica, Cuba, Ecuador, Mexico, Peru, and Trinidad and Tobago.

technologies and research platforms. International cooperation aids local researchers to overcome such bottlenecks (Soler, 2021). Argentina, for example, has developed a national strategy to strengthen marine research capabilities, allowing for better coordination and optimization of resources, the "Pampa Azul" initiative. However, seven years after its launching, economic instability jeopardized investments, despite international commitments. One current opportunity is the All-Atlantic Ocean Research Alliance, a South to North multilateral scientific cooperation open to countries in the region (Polejack et al., 2021). It aligns research priorities, infrastructure, and budget, to overcome the knowledge gaps in the Atlantic, informing decisions for improved societal benefit. As a result, the Alliance fosters marine technology transfer and balanced knowledge co-production.

Use of regional knowledge is hindered in world assessments for several reasons. Despite budget constraints, political and economic instability, LAC researchers produce a wealth of knowledge that often faces intra-academic barriers, such as language (Angulo et al., 2021). Knowledge relevant to local systems are often published in languages other than English or outside of mainstream Journals receiving less attention by peers and thus becoming invisible to global assessments like SROCC. Therefore, local researchers are again critical to make such knowledge visible to global reporting processes. Regional ILK needs to permeate more effectively into global assessments like IPCC and IPBES reports to complement classic scientific information. At the same time, the results/findings from such reports must return to local communities in a language that both society and policymakers understand and relate to, so that the uptake of such knowledge is enhanced. Science cannot be detached from local realities, even if the final message pertains to global effects. Translation of scientific knowledge to local languages is certainly essential to allow for a more equitable extraction of the information from these assessments making calibrated language and the whole process more palatable to the general public since its information is designed to provide evidence, agreement and communicate uncertainties (Mastrandrea et al., 2010) based on peer-reviewed research. By adjusting the language, it allows the information and its flow to be more inclusive and receptive to diversity, particularly when interculturality is taken into account. Although many perceive the region as sharing similar languages, geographies and cultures, reality shows a huge diversity of languages and cultures but also values and beliefs. Latin America and the Caribbean are as diverse and wide as the geographic breadth and ecosystems/biomes described in-between. Thus, climate change perception of threats needs to account for local realities and require larger representation of specific groups, knowledge and traditions at different assessment processes.

Climate knowledge production is also dependent on multiscale observing systems to produce accurate scenarios and long-term predictions. Nevertheless, despite existing initiatives, there are still considerable capacity and data gaps (Malone et al., 2010; Foltz et al., 2019; Smith et al., 2019; Speich et al., 2019; IOC-UNESCO, 2020) due to insufficient observations. These gaps are particularly critical in coastal Africa, South America, the Caribbean, Southeast Asia, and Small Island Developing States, where development pressure and high social vulnerability hamper ocean and climate sustained observations. These areas should represent high monitoring priorities and efforts.

Knowledge Accessibility

Societal engagement is influential in science uptake to inform decisions by pressing governments to act, as well as by using scientific information to transform behavioral patterns and foster climate and ocean literacy, and social innovation. Consequently, inclusion and equity require accessible language and capacity development. In socio-ecological systems, where scientific uncertainty and societal stakes are high, values tend to be in dispute and decisions are urgent. The Post-Normal Science framework (PNS, Funtowicz and Ravetz, 1993) proposes a multi-stakeholder engagement in the decision making, jointly considering the risks and opportunities to act. We need to address social vulnerabilities at the local level to enhance and sustain the engagement of LA in the green/blue economy. Thus, PNS could be an adequate framework for developing SROCC's recommendations further.

Scientific knowledge used to be restricted to academic groups and publications and discussed within invisible schools (Sieber, 1991). Recently, the Open Science Movement has attempted to make this knowledge available to all (Aspesi and Brand, 2020). Open Science is about making scientific research and data freely accessible, but should also mean dialoguing with society, while embracing ILK in support of better-informed decision making (Oliver and Cairney, 2019; Safford and Brown, 2019).

Broad stakeholder engagement (affected communities, indigenous peoples, local and regional representatives, policy makers, managers, interest groups and organizations) has the potential to combine and use relevant knowledge (Obermeister, 2017) and balance the disproportionate influence that economically advantaged groups have in most scientific assessments (Castree et al., 2014). The formal process of IPCC assessments follows predetermined formats and standards[7,8], uses specific calibrated language and approaches unfamiliar to many scientists and policy makers in LA. Locally, there is little interaction and support by IPCC focal points to promote learning-oriented methodologies, familiarity with the language and experience to address the IPCC process, hampering regional/local participation. Although the recognition and use of ILK is expanding in peer-reviewed research (Savo et al., 2016; Abram et al., 2019) thus providing information and responses to guide and inform policy with different perspectives (Huntington, 2011; Nakashima et al., 2012; Lavrillier and Gabyshev, 2018), most global assessments have not yet incorporated ILK information (Obermeister, 2017) thus limiting the potential of local adaptation response (Ford et al., 2016).

Science diplomacy, the interrelation between research and international relations, can reduce inequalities and bridge communities by aiding in the implementation of international

[7]https://www.ipcc.ch/documentation/procedures/

[8]https://www.ipcc.ch/site/assets/uploads/2018/09/ipcc-principles-appendix-a-final.pdf

provisions aimed at leveraging scientific capabilities in LAC (Ruffini, 2018; Salpin et al., 2018; Polejack and Coelho, 2021). By incorporating scientific literature in other languages, other sources of knowledge, and regional input, global assessments like SROCC reduce most of its imbalance. The opportunity presented by the UN Decade of Ocean Science (Ryabinin et al., 2019; Polejack, 2021), particularly through the Ocean Literacy movement seek creative ways to bridge science, policy, diplomacy and society (Santoro et al., 2017; Borja et al., 2020).

Knowledge Impact to Policy Change

Climate change adaptation and mitigation requires coherence of global, national, and local levels of governance, a challenge to the integration of political and administrative systems. There is a void between international treaties, national regulations, and local implementation due to the lack of broad stakeholder participation in the formulation of these policies, undermining their adequacy (Keskitalo et al., 2016). The development of effective responses involves societal adjustment and modification of current behavior provoking such changes.

Scientific advice is playing an increasing role in policy and decision-making (Gluckman, 2016a). Governments require scientific evidence in a wide range of situations (e.g., Gluckman, 2016b), but there is still the need to respect the different imperatives in science and in policy, so better-informed decisions are made, and research is promoted and sustained in the long-term (Parkhurst, 2016).

Interculturality matters to LAC (UNDP, 2021) and has been recognized as an important regional aspect that defines local identity as reflected in a few NDCs. Thus, as scientists, we must incorporate the local social, cultural, and political forces to seek mutual understanding and cooperation to also find solutions to climate change adaptation. Local institutional and policymaking landscapes are determinant of how scientific evidence is perceived and used in the decision-making process, mostly because these decisions consider a wide range of factors that are grounded on local realities, including social values and beliefs (Cairney, 2016) and traditional and local knowledge, reflected in the interculturality aspects brought by a few NDCs. Latin America has a diversity of political systems that produce and apply scientific evidence in a variety of ways, deriving from national and subnational realities that often challenge the Western-democratic perspective of the use of evidence, so dominant in global reporting exercises (Parkhurst, 2016). Thus, standard global solutions can become locally irrelevant and there is a need to consider these realities when co-designing fit-for-purpose local solutions. In this sense, local actors (scientists, the public and stakeholders) are better equipped to act as knowledge brokers within their local social-political contexts.

Importance of Long-Term Monitoring for Decision-Making

Long-term observations inform society about change rates in ocean warming, sea-level rise, acidification, and deoxygenation (Breitburg et al., 2018; Bourlès et al., 2019; Turk et al., 2019), including coastal areas where the effects on ecosystems and ecosystem services are often associated with social vulnerability, highly affecting society (IPCC, 2019b). Detection of climate change in coastal regions is difficult because of their natural variability, requesting long-term ocean observing systems (Duarte et al., 2013; Turk et al., 2019). Globally coordinated ocean observing systems provide the information needed to support climate prediction on different timescales (e.g., Sloyan et al., 2019). However, many existing records are still short to detect anthropogenic change, and some regions remain undersampled (e.g., deep-sea, shelves). Southern Hemisphere temperate, subpolar and polar latitudes are among the least studied areas of the planet, which represents a serious gap to decrease the uncertainty of global models predicting future climate scenarios (Meredith et al., 2019). Long-term data is essential to measure changes to ecological and environmental conditions, but also the outcome of policies and human behavioral changes (Pecl et al., 2017). Thus, long-term ocean observatories in LAC, combining environmental data (such as Essential Ocean Variables-EOVs) with social sciences and traditional knowledge need to be developed and implemented (Abram et al., 2019; Fennel et al., 2019).

At the heart of climate change research is the requirement of sustained observations with time series frequent and long enough to develop baselines and climatologies. Baselines are compared with anomalies, changes in phenology, trends or changes in populations, and spatial distribution. Time series enables us to characterize variability, reduce uncertainty, and increase forecast and prediction which can guide the outcome of policies and human behavioral and environmental change. Bio-Environmental baselines and time series represent global trends and local pressures that can be evaluated against natural variation for policy and decision-making at many levels (Muelbert et al., 2019). Integrative scenarios, combining environmental, socioeconomic and health sciences, such as the Nexus method (Howells et al., 2013), has been successfully applied to climate and fisheries in the Humboldt Current System (Garteizgogeascoa et al., 2020), in the assessment of climate vulnerability in Brazil (Araujo et al., 2019), and in the International Long-Term Ecological Research (LTER) programs described in Muelbert et al. (2019) and detailed for LAC in **Table S2**.

Consequently, better government climate-related decisions are likely to occur when decision-makers are exposed to climate scenarios and environmental indicators with dynamic outputs, even in face of models' limitations and potential risks of being misused to support biased political statements (Saltelli et al., 2020). According to Haasnoot et al. (2015), scenarios lead to increased awareness of when and which adaptation policies should be applied.

CONCLUSIONS

Despite the efforts to disseminate, warn and engage as many nations as possible in a global effort to reverse the course of climate change, high inequality and low economic growth in several regions are hampering the overall understanding, integration, and engagement to mitigate climate effects, thus perpetuating regional heterogeneity (UNDP, 2021). The goals and specific objectives of climate change strategies around the world tend to reflect a global agenda that, at least for LA, are often detached from national/regional vulnerabilities and contexts which in part respond to delayed actions. It needs to change.

In order to reduce knowledge gaps in LA, there is a need to secure investment in long-term observations and to promote capacities, which will also raise the accuracy of models and predictions. Sustainable research funding shall provide local and regionally oriented information and advice. Moreover, successful initiatives like Pampa Azul, AtlantOS, the All-Atlantic Ocean Research Alliance, Rede Clima, Acceso Libre a la Información Científica—ALICIA, the National Repository in Mexico and the Cartagena Convention (**Table S1**) reflect State policies trying to overcome bottlenecks in LA. The interruption of such policies jeopardizes future investments and continuity of climate action mitigation.

How would Latin America engage in climate action globally while maintaining its identity and structure of interconnected social, economic, and ecological systems? It is imperative to develop specific national-institutional capacities and public awareness to support and advance a long-term process with a more diverse and multi knowledgeable approach embracing local cultures, language, and broader participation of local communities (**Figure 1**). Despite political and economic limitations, the region must be integrated not only from a commercial perspective of goods and services (i.e., Mercosur) but also from an environmental standpoint to implement its strategies against irreversible climate change. A few organizations in the region could facilitate this coordination and strengthen the participation of LAC representatives in global reporting assessments such as SROCC. The InterAmerican Institute for Global Change (IAI—Instituto InterAmericano para la Investigación del Cambio Global) is a regional intergovernmental organization that promotes interdisciplinary scientific research and capacity building, informing local and regional decision-makers about important issues of global change. Although the IAI has mechanisms in place to provide scientific evidence for the improvement of its Parties' public policies, it is essentially intergovernmental, i.e., triggered by diplomatic negotiations that depend upon national mechanisms of integration with other stakeholders. The Economic Commission for Latin America and the Caribbean (ECLAC) is also another important intergovernmental organization in the region that could enhance the coordination in climate change responses, significantly

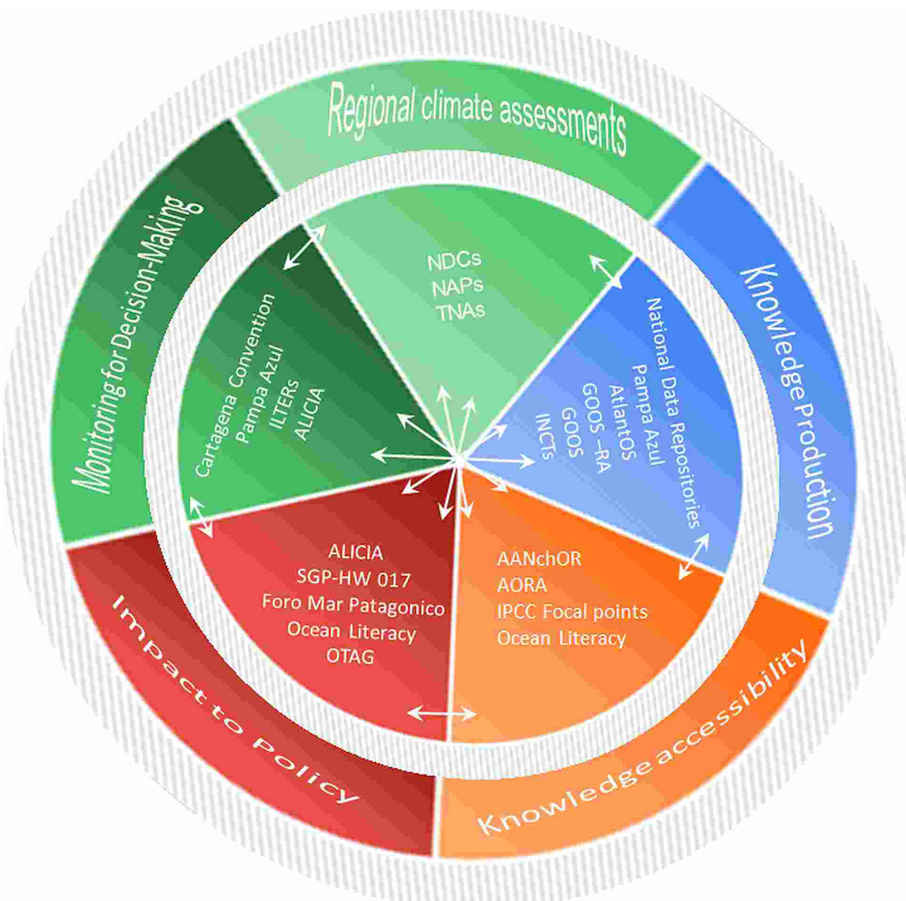

FIGURE 1 | Examples of current initiatives that can narrow the gaps and explore opportunities for the uptake of ocean and climate science to political decision making in LA across five key categories: (i). climate assessment information and regional policies, (ii). knowledge production, (iii). knowledge accessibility, (iv). knowledge impact to policy, and (v). long-term monitoring for decision-making. (Note see **Table S1** for additional information and references).

contributing to regional knowledge production and public policies, while promoting the transition to environmentally sustainable and low carbon economies (UNDP, 2021). However, both IAI and ECLAC apply similar diplomatic processes as the IPCC and the UNFCCC, with little synergy with local stakeholders. Moreover, neither have a climate (not to mention ocean) focus and not all countries in the region are Parties to those organizations. Therefore, while we recognize that regional organizations can aid in bridging global, regional and national perspectives based in science, we advocate that local researchers can act as knowledge brokers and should be empowered, encouraged, rewarded and better included in global climate-ocean scientific assessments.

Addressing climate change entails modifying the status quo facing resistance from influential groups in society that interfere with the development of local climate change policies (Meadowcroft, 2009). In this perspective, we advocate that the uptake of SROCC findings in LA policies can be enhanced by: (a) embracing local realities and knowledge purveyors; (b) empowering locals to both inform local knowledge to global assessments and adapt those findings to local realities; (c) enhancing regional research capabilities; and (d) securing ocean observations for the long run. The adoption and incorporation of SROCC's recommendations into NDCs depend strongly on the local reality which is dictated by the relationship between adaptation-related processes (social vulnerability, low growth, as well as high contrasts and inequalities) and political pressures.

DATA AVAILABILITY STATEMENT

The original contributions presented in the study are included in the article/**Supplementary Material**, further inquiries can be directed to the corresponding author.

AUTHOR CONTRIBUTIONS

MMCM led the process with all co-authors (MC, LCC, MNL, AP, ACP-P, and ER-A) who have substantially contributed to the document, its revision, reading and approving the submitted version.

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SUPPLEMENTARY MATERIAL

The Supplementary Material for this article can be found online at: https://www.frontiersin.org/articles/10.3389/fclim.2021.748344/full#supplementary-material

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Siberian Ecosystems as Drivers of Cryospheric Climate Feedbacks in the Terrestrial Arctic

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Climate warming is altering the persistence, timing, and distribution of permafrost and snow cover across the terrestrial northern hemisphere. These cryospheric changes have numerous consequences, not least of which are positive climate feedbacks associated with lowered albedo related to declining snow cover, and greenhouse gas emissions from permafrost thaw. Given the large land areas affected, these feedbacks have the potential to impact climate on a global scale. Understanding the magnitudes and rates of changes in permafrost and snow cover is therefore integral for process understanding and quantification of climate change. However, while permafrost and snow cover are largely controlled by climate, their distributions and climate impacts are influenced by numerous interrelated ecosystem processes that also respond to climate and are highly heterogeneous in space and time. In this perspective we highlight ongoing and emerging changes in ecosystem processes that mediate how permafrost and snow cover interact with climate. We focus on larch forests in northeastern Siberia, which are expansive, ecologically unique, and studied less than other Arctic and subarctic regions. Emerging fire regime changes coupled with high ground ice have the potential to foster rapid regional changes in vegetation and permafrost thaw, with important climate feedback implications.

Keywords: Siberia, larch, permafrost, wildfire, snow—vegetation interactions, ecosystems, arctic and boreal ecosystems, Arctic

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INTRODUCTION

Amplified climate warming across the Arctic is causing widespread terrestrial cryospheric change in the form of altered snow cover dynamics and permafrost thaw, both with the potential to act as globally important climate feedbacks (Flanner et al., 2011; Schuur et al., 2015; Meredith et al., 2019). Reduced albedo associated with declines in snow cover duration and increased snow masking by vegetation canopies protruding above the snowpack acts as a positive climate feedback (Bonan et al., 1992; Chapin, 2005). Permafrost thaw makes previously inaccessible organic carbon available for microbial decomposition and subsequent release to the atmosphere as greenhouse gases, also acting as a positive feedback to climate warming (Schuur et al., 2015). These processes are not independent; increases in snow depth insulate permafrost soils from cold winter air temperatures, further increasing greenhouse gas efflux and the likelihood of thaw (Stieglitz et al., 2003; Anisimov and Zimov, 2020). Consequently, concurrent changes in snow cover and permafrost thaw across the terrestrial Arctic and subarctic will dictate the strength of terrestrial cryospheric climate feedbacks. However, quantifying these changes remains a key challenge in climate research because they are affected by spatially heterogeneous interactions between climatic, ecological, and geomorphic conditions, as well as disturbances such as fire (Loranty et al., 2018).

Among ecological factors, vegetation is a key determinant of cryospheric feedback strengths because of the role it plays in surface energy partitioning (Loranty et al., 2018). Variability in vegetation canopy cover, height, and leaf habit within and between vegetation functional types at local and regional scales drives differences in land surface albedo that are important in the context of snow-albedo-feedback magnitudes (Sturm et al., 2005; Loranty et al., 2014). Similarly, vegetation canopy influences on snow redistribution impact the timing and duration of melt-out (Pomeroy et al., 2006; Marsh et al., 2010; Sweet et al., 2014) and soil thermal dynamics (Kropp et al., 2020). Without snow cover, vegetation canopy shading may reduce permafrost soil temperatures, while plant water use and soil properties interact to affect ground heat flux (Blok et al., 2010; Loranty et al., 2018). Taken together, these processes highlight the important fact that an accurate predictive understanding of terrestrial cryospheric change across the northern hemisphere requires detailed knowledge of ongoing vegetation change. In particular, changes that involve transitions from prostrate to erect shrub growth forms, transitions in leaf habit (e.g., deciduous vs. evergreen), or substantial changes in shrub and forest canopy cover are especially important.

Vegetation structure and distribution are changing in direct response to warming temperatures, with the most prominent example being shrub expansion into graminoid dominated tundra ecosystems (Tape et al., 2006; Forbes et al., 2010; Frost and Epstein, 2014). Here, the associated reduced albedo (Loranty et al., 2011), snow trapping by canopies (Sturm et al., 2001), and increased ground temperatures (Myers-Smith and Hik, 2013; Kropp et al., 2020) are well documented. Wildfire disturbance is also an important driver of vegetation change, with altered post-fire succession as a consequence of climate-driven fire regime shifts in boreal North America forming the primary basis of current knowledge (Johnstone et al., 2010; Alexander and Mack, 2016). Associated post-fire albedo recovery (Amiro et al., 2006; Randerson et al., 2006), albedo associated with deciduous dominance (Beck et al., 2011), and elevated ground temperature (Fisher et al., 2016; Way and Lapalme, 2021) are among the key consequences. Notably, however, there has been less work focused on boreal canopy-snow interactions, and previous fire focused work occurs mostly in the discontinuous and sporadic permafrost zones where permafrost may not be present. Conspicuously absent from this emerging body of research are analyses focused on ecosystem-cryosphere interactions in the Siberian Arctic and subarctic (Metcalfe et al., 2018), despite a unique combination of ecological and cryospheric conditions and extensive areal extent that makes it regionally important. In this perspective, we highlight and synthesize recent research and emerging evidence illustrating the potential for ecosystem processes to drive rapid and widespread cryospheric change across central and eastern Siberia, with particular focus on larch (*Larix* spp.) dominated forests.

THE SIBERIAN CLIMATIC AND ECOLOGICAL SETTING

Eastern Siberia is characterized by winters with extreme low air temperatures with thin to moderate snow accumulation that reaches maximum depths of 20–100 cm between January and March (Grippa et al., 2004; Zhong et al., 2018; Mortimer et al., 2020). Maximum snow depth increased throughout much of Siberia over recent decades (Bulygina et al., 2009). Snow accumulation typically begins around October, and snow extent is largely diminished by June (Grippa et al., 2004). The duration of total days of snow cover is decreasing in northern and central regions (Bulygina et al., 2009; Bormann et al., 2018) coinciding with below average May and June snow cover extent in recent decades (Mudryk et al., 2017, 2020). On the ecosystem and landscape scales, patterns of snow depth and timing of snowmelt can be varied, influenced by localized factors including topography, vegetation canopy characteristics, and tree density (Sturm et al., 2001; Todt et al., 2018).

Continuous permafrost occurs at subarctic latitudes (~60°N) in central and eastern Siberia, owing to the harsh continental climate. Thick, carbon and ice-rich permafrost deposits (termed Yedoma) are prevalent in Northern Siberia, particularly in lowlands and foothills (Grosse, 2013; Strauss et al., 2017). The high ice content (~80% by volume) of Yedoma deposits means they are highly vulnerable to temperature-induced permafrost degradation, and thermokarst lakes are prevalent throughout the region (Olefeldt et al., 2016; Nitze et al., 2018). Permafrost thaw has also been documented in the more mountainous areas of Siberia, where hill slope thermokarst features such as retrogressive thaw slumps are common (Olefeldt et al., 2016; Nitze et al., 2018).

Siberia is characterized primarily by boreal forests and wetlands with a transition to tundra ecosystems a few degrees of latitude north of the Arctic circle. Considerable landscape heterogeneity is driven by local variation in topoedaphic and hydrologic conditions (Sulla-Menashe et al., 2011). Tundra vegetation communities are comparable to those found in North America, with similar climatic and geomorphic controls on distribution (Walker et al., 2005). Relative to North America, the forest-tundra ecotone generally occurs further north and is more diffuse in Siberia (Ranson et al., 2011) owing to differences in geomorphology and glacial history. Siberian boreal forests differ in several key ways from those in North America. Eastern Siberian boreal forests are dominated by larch (*Larix* spp.), a deciduous needle-leaf tree. Larch forests tolerate cold soil conditions, and cover more than 3 M km^2 with in the permafrost zone (Loranty et al., 2016). By contrast, boreal forests in North America and western Siberia are composed of evergreen conifers and deciduous broadleaf trees and are underlain by sporadic or isolated permafrost.

The deciduous, needle-leaf larch (*Larix* spp.) forests that cover vast expanses of Siberia's boreal biome require periodic

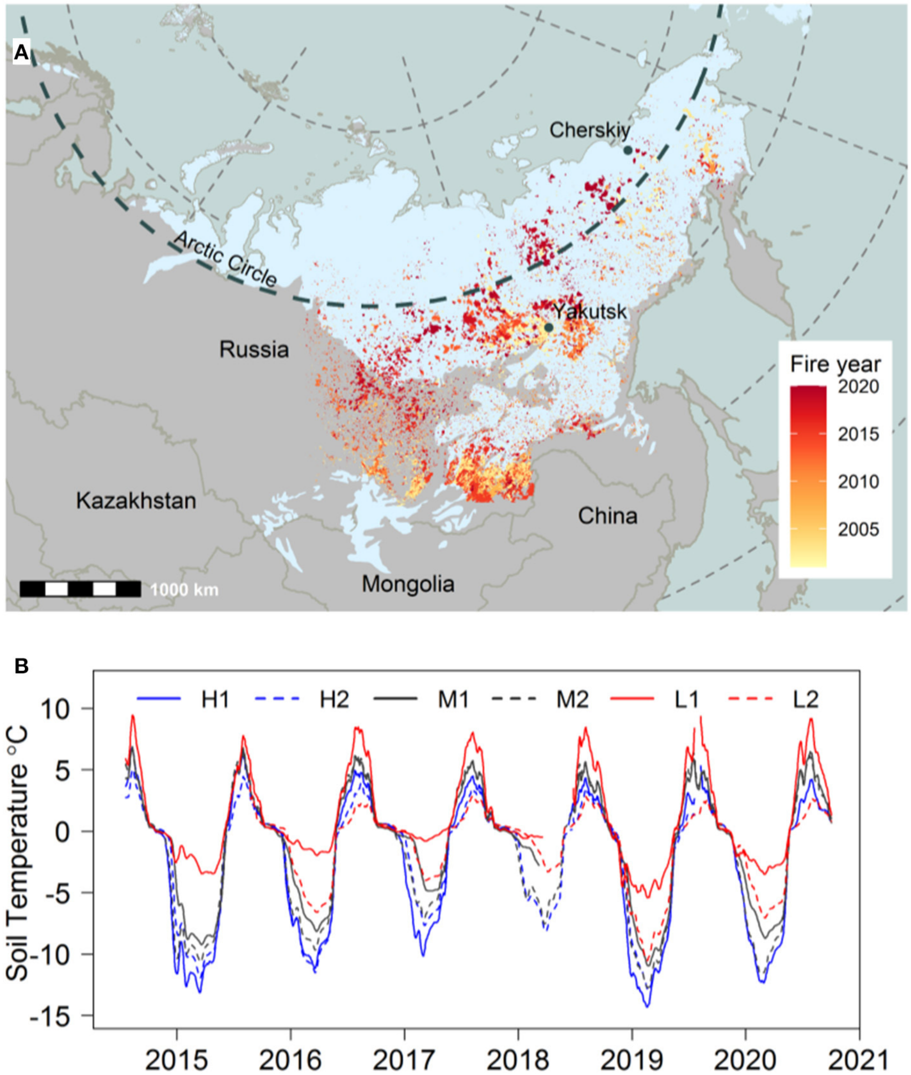

FIGURE 1 | Area burned **(A)** in eastern Siberia from 2001–2020 derived from Landsat data using Google Earth Engine (Talucci et al., 2021) with areas of continuous permafrost delineated in light blue. Note more area has burned (in red) in recent years above the Arctic circle (gray dashed line). Soil temperature **(B)** across a forest density gradient associated with post-fire regrowth within a burn perimeter near Cherskiy where H1 and H2 represent high-density stands (> 75% canopy cover), M1 and M2 represent medium density stands (75% > canopy cover > 25%), and L1 and L2 represent low density stands (<25% canopy cover) (Loranty and Alexander, 2021).

fire disturbances to perpetuate their long-term persistence in this region (Kharuk et al., 2011). Fires, ignited by lightning or humans during the warm and relatively dry growing season (Kharuk et al., 2021), occur every 80–350 years, with fire return intervals increasing with latitude (Kharuk et al., 2011; Berner et al., 2012). Fires are typically surface fires (>90%) that creep along the forest floor, fueled by consumption of the thick soil organic layer (SOL) that accumulates during the fire free intervals (Kharuk et al., 2021). Fire removal of the SOL exposes high-quality mineral soil seed beds that promote seed germination (Sofronov and Volokitina, 2010; Alexander et al., 2018). Fire also consumes shrubs and other vegetation, decreasing competition, and improving conditions for early seedling establishment (Kharuk et al., 2011, 2021). Without fire to prepare the seedbed and remove competition, larch recruitment is often limited to microsites provided by tip-up mounds or other small-scale disturbances (Shirota et al., 2006).

Fire activity varies across Siberia, and much of what is known comes from research in western and central Siberia with a limited understanding of fire activity in eastern Siberia. Across Siberia, the area burned and the number of large fire seasons (i.e., over 50% area burned) has increased between 1995 and 2005 (Soja et al., 2007), but the full extent of area burned is often underestimated by coarse resolution satellites (Shvidenko

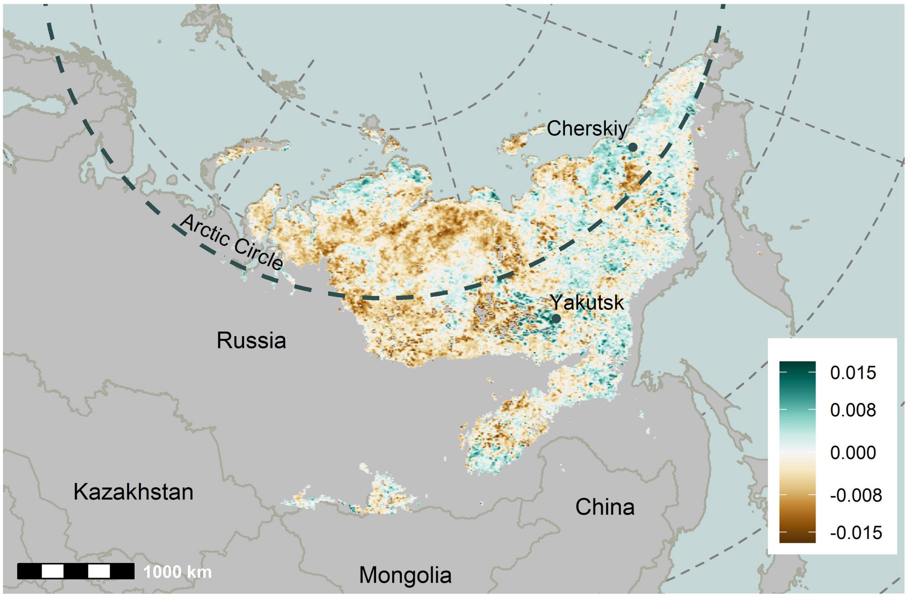

FIGURE 2 | Trends in albedo change (April–September of 2000-2019) attributed to surface water change derived using MODIS data. The standard MODIS albedo product was used to characterize albedo trends using linear regression, and the surface water change was quantified using the Superfine Water Index calculated from MODIS data. A generalized additive model (GAM) was used to attribute albedo trends to surface water change. Adapted from Webb et al. (2021).

et al., 2011; Berner et al., 2012) e.g., MODIS ∼500 m; (Shvidenko et al., 2011; Berner et al., 2012). It is predicted that greater combustion losses and permafrost thaw during more extreme fire seasons will increase carbon emissions (Soja et al., 2004). In northwestern Siberia, warmer and drier conditions (Soja et al., 2007; Ponomarev et al., 2016) as well as human activity (Sizov et al., 2021) have contributed to increased fire activity. In northeastern Siberia fire regimes are less well understood (Berner et al., 2012) because long-term ground-based records are limited and difficult to access (Soja et al., 2007). Improved quantification of regional fire dynamics is crucial to identifying fire-induced vegetation shifts and altered soil conditions (Holloway et al., 2020) that act as catalysts for permafrost degradation (Loranty et al., 2016), and contribute further to ongoing climate feedbacks.

Larch forest recovery after fire requires a co-occurring suite of conditions. Fires often damage larch roots (Kharuk et al., 2021), which are mostly restricted to the shallow, seasonally-thawed ground layer above the permafrost (Abaimov et al., 1998). Consequently, many mature, seed-bearing larch die soon after fire, resulting in stand-replacing fires. Thus, a critical primary filter for larch forest recovery after fire is seed availability, which depends on several factors. Larch annually produce small, wind-dispersed seeds that can travel ∼100 m (Cai et al., 2018), with masting events (large seed crops) every 3–5 years (Abaimov, 2010). As such, small seed crops in non-mast years can limit post-fire larch recovery despite fire's creation of appropriate seedbed conditions. However, if the seed crop is substantial, and fires do not propagate into the larch canopy, already ripe seeds on fire-killed trees can provide seed sources for forest recovery (Kharuk et al., 2021). If canopy fires damage seeds, recovery depends on seeds from nearby unburned forests or islands of unburned trees. Thus, seed availability may be low within the interior of large burn patches whose size exceeds that of larch seed dispersal distances (Cai et al., 2013), especially in low seed crop years, unless residual seed crops persist within the burn (Cai et al., 2018). Recruitment failure may occur in areas with ice-rich yedoma permafrost where charring and vegetation removal triggers ground subsidence and pooling of water on the ground surface inhibits larch seed germination (Alexander et al., 2018). Ultimately, low seed availability or failure to germinate after fire can result in larch recruitment failure and forest transition to alternative vegetation cover types, including shrublands and grasslands (Cai et al., 2013, 2018; Chu et al., 2017).

DISCUSSION OF EMERGING ECOSYSTEM CHANGES WITH IMPLICATIONS FOR CRYOSPHERIC CLIMATE FEEDBACKS

Vegetation cover is an important determinant of the snow albedo feedback (Thackeray et al., 2019). As with elsewhere in the Arctic (Beck and Goetz, 2011), tree and shrub cover are increasing across Siberia in response to rising temperatures (Frost and Epstein, 2014; Shevtsova et al., 2020, 2021). The prevalence of canopy infilling in larch forests (Shevtsova et al., 2020) combined with relatively low canopy cover (Loranty et al., 2014) and fragmented nature of the forest-tundra ecotone (Ranson et al., 2011) in the region create a potential for changes in forest cover over large areas. These vegetation changes will act as a positive climate feedback when increased canopy cover masks the underlying snowpack thereby reducing albedo. However, when vegetation masks snow the difference between snow-on and snow-off albedo is lower, and the strength of climate feedbacks associated with changes in the duration of seasonal snow cover is reduced (Euskirchen et al., 2016; Potter et al., 2020). Accurate representation of ongoing vegetation change is therefore integral to observational quantification and predictive modeling of the snow albedo feedback strength.

Fire is an important driver of vegetation change, and though it is common in the region, the past several years have been exceptional both in terms of total area burned, as well as the proportion of area burned above the Arctic Circle. In northeastern Siberia, the 2020 fire season alone constitutes nearly 30% of the cumulative area burned over the past two decades. During the 2019 and 2020 fire seasons the area burned in the Arctic (3.4 and 5.4 million hectares, respectively) was much greater than the average over the previous two decades (0.7 million hectares) (**Figure 1A**; Talucci, in review). Additionally, recent albedo trends suggest a fire regime shift may be underway (Webb et al., 2021). If these fire regime shifts are sustained, the potential impacts on forest cover associated with recruitment dynamics (Alexander et al., 2012, 2018; Paulson et al., 2021) have important implications for cryospheric climate feedbacks. The extent to which increases in winter albedo immediately postfire (Chen et al., 2018; Stuenzi and Schaepman-Strub, 2020) are reversed during post-fire succession will depend on the density of forest regrowth (Talucci et al., 2020; Paulson et al., 2021; Walker et al., 2021). Similarly, forest impacts on surface energy exchange exert a series of interrelated controls on permafrost and snow thermal dynamics (Loranty et al., 2018; Stuenzi et al., 2021). Variation in forest recovery within a single fire perimeter near Cherskiy has resulted in even-aged stands with canopy cover ranging from ∼10–95% (Paulson et al., 2021), and winter albedos of ∼0.3 and ∼0.7 at representative high- and low- density stands (Kropp and Loranty, 2018; Kropp et al., 2019). Corresponding to post-fire tree density, summer and winter soil temperatures are warmer at sites with lower tree cover (**Figure 1B**) as a result of lower canopy shading and insulation from organic soils, and snow redistribution, respectively. Fire-induced vegetation change will exert strong control on albedo and permafrost climate feedbacks if this type of variation in post-fire larch regrowth becomes more common. Consequently, accurate model predictions of these cryospheric climate feedbacks requires improved observational understanding of fire impacts on vegetation-snow-permafrost interactions in the region.

Changes in snow cover, and snow-vegetation interactions, are likely to influence the distribution and rates of permafrost thaw. Snow is an effective insulator that decouples permafrost soils from frigid winter air temperatures (Jorgenson et al., 2010; Park et al., 2015), and numerous manipulation experiments have illustrated that increases in snow depth can lead to permafrost thaw and increased greenhouse gas emissions (Walker et al., 1999; Natali et al., 2014; Webb et al., 2016). For example, in northeastern Siberia near Cherskiy, snow depths were nearly double the long-term average in the winters beginning in 2016 and 2017, resulting in exceptionally deep seasonally thawed active layers that did not fully re-freeze in subsequent winters (Anisimov and Zimov, 2020). While these findings are consistent with model projections (Räisänen, 2008; Mankin and Diffenbaugh, 2015), and the relationship between snow cover and permafrost soil temperature is not unique to Siberia (Lawrence and Slater, 2010), the high ground ice and carbon content of Siberian permafrost mean that ground thaw in the region will be irreversible once started and lead to high carbon emissions (Strauss et al., 2017). The observations reported by Anisimov and Zimov (2020) highlight the following important points regarding the impacts of snow cover on permafrost climate feedbacks. First, changes in snow cover can initiate permafrost thaw even in regions such as northeastern Siberia where air and permafrost temperatures are low (Anisimov and Zimov, 2020). Additionally, the relative impact of increased snow depth on permafrost varies with vegetation and disturbance history (Shur and Jorgenson, 2007). This highlights the importance of vegetation snow interactions, which are well documented at the local scale (Essery and Pomeroy, 2004), and appear to be important yet less well understood at the pan-Arctic scale (Kropp et al., 2020). Lastly, Anisimov and Zimov (2020) noted the greatest impacts of increased snow cover on permafrost in recently burned locations devoid of vegetation and organic soils, which is in line with observations of an inverse relationship between seasonally thawed active layer depth and time since fire for forests in the region (Webb et al., 2017; Alexander et al., 2018; Kharuk et al., 2021). Taken together, this evidence indicates that vegetation-snow interactions are of crucial importance for albedo and permafrost related climate feedbacks. The ability of fire to foster rapid vegetation change over large areas in relatively short time spans means that changing fire regimes may exert very strong influences on regional cryospheric feedbacks in the coming decades.

Wildfire, and climate warming more generally, thaw permafrost and cause ground subsidence or thermokarst formation (Jones et al., 2015; Yanagiya and Furuya, 2020). Thermokarst features, especially lakes, are common in Siberia, and surface water extent (e.g., lake size, river discharge, and moisture stored in the active layer) is changing across the region—increasing in some areas and decreasing in others—likely due to climate-change induced shifts in precipitation and permafrost thaw (Iijima et al., 2010; Karlsson et al., 2012; Vey et al., 2013; Boike et al., 2016; Nitze et al., 2017; Wang et al., 2021). Changes in the areal extent of surface waters have important implications for localized albedo-mediated heat transfer (that is, the presence of water can increase near surface temperatures by as much as 10°C; (Jorgenson et al., 2010) as well as regional warming and cooling through albedo-induced changes in top-of-the atmosphere radiative forcing (Webb et al., 2021). Overall, changes in Siberian surface water over the past two decades have led to regional albedo-induced warming (Webb et al., 2021), although the pattern is heterogeneous across the region (**Figure 2**). Thermokarst lakes are also CH_4 emission hotspots (Heslop et al., 2020) that effectively increase global warming potential of greenhouse gas emissions related to permafrost thaw.

CONCLUSIONS

Eastern Siberia contains a unique combination of fire-prone larch forests and ice-rich yedoma permafrost. The ability of fire to alter vegetation and permafrost conditions over relatively short time scales has the potential to influence climate feedbacks associated with albedo change and permafrost thaw. However, our understanding of ongoing change in the region is lacking relative to other areas of the Arctic, thus highlighting a need for additional work in the region. Specifically, there is a crucial need to understand how emerging changes in the fire regime are likely to impact post-fire vegetation recovery, and how this in turn will affect permafrost resilience. Variation of vegetation-snow interactions, and associated impacts on permafrost stability also requires closer examination. The latter is particularly important in the context of surface water induced albedo trends. Accurately predicting the timing and magnitude of the cryospheric climate feedbacks arising from terrestrial ecosystem change in the region requires improved understanding of key processes achieved through continued observational and model investigation. For example, a recent study in northern Siberia conducted using a model with explicit representation of ice-rich permafrost dynamics, informed with field observations, predicted a much greater degree of permafrost thaw than previous projections that lacked these process representations (Nitzbon et al., 2020). Similar research utilizing observations and models to investigate interactions between fire, vegetation, and permafrost will help improve forecasts of ecological change in the region and identify key processes necessary to help constrain feedback processes in land surface and climate models. It is also important to note that a great deal of research on this region has been published in the Russian literature, and so translation of Russian papers and data offers a means to improve understanding of the region. Additionally, eastern Siberia includes large indigenous populations, who can provide additional insight to cryospheric change given their reliance on cryospheric resources for a range of basic needs (Fedorov, 2019), and through their long-term landscape-based perspectives that may be missed by western scientific research (Crate et al., 2017).

DATA AVAILABILITY STATEMENT

The datasets shown in Figure 1 can be found at the Arctic Data Center (**Figure 1A**; https://doi.org/10.18739/A2N87311N, **Figure 1B**; https://doi.org/10.18739/A24B2X59C). Data from **Figure 2** are publicly available MODIS data as described by Webb et al. (2021).

AUTHOR CONTRIBUTIONS

ML, AT, and EW prepared the figures. All authors contributed to conception and design of the paper, and wrote sections of the manuscript, contributed to manuscript revisions, read, and approved the submitted version of the manuscript.

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Regime Shifts in Glacier and Ice Sheet Response to Climate Change: Examples From the Northern Hemisphere

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Glaciers and ice sheets are experiencing dramatic changes in response to recent climate change. This is true in both mountain and polar regions, where the extreme sensitivity of the cryosphere to warming temperatures may be exacerbated by amplification of global climate change. For glaciers and ice sheets, this sensitivity is due to a number of non-linear and threshold processes within glacier mass balance and glacier dynamics. Some of this is simply tied to the freezing point of water; snow and ice are no longer viable above 0°C, so a gradual warming that crosses this threshold triggers the onset of melting or gives rise to an abrupt regime shift between snowfall and rainfall. Other non-linear, temperature-dependent processes are more subtle, such as the evolution from polythermal to temperate ice, which supports faster ice flow, a shift from meltwater retention to runoff in temperate or ice-rich (i.e., heavily melt-affected) firn, and transitions from sublimation to melting under warmer and more humid atmospheric conditions. As melt seasons lengthen, there is also a longer snow-free season and an expansion of glacier ablation area, with the increased exposure of low-albedo ice non-linearly increasing melt rates and meltwater runoff. This can be accentuated by increased concentration of particulate matter associated with algal activity, dust loading from adjacent deglaciated terrain, and deposition of impurities from industrial and wildfire activity. The loss of ice and darkening of glaciers represent an effective transition from white to grey in the world's mountain regions. This article discusses these transitions and regime shifts in the context of challenges to model and project glacier and ice sheet response to climate change.

Keywords: non-linear feedbacks, tipping point, albedo, meltwater retention, glacier mass balance, mountain glaciers

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INTRODUCTION

The IPCC (2019) special report on oceans and the cryosphere in a changing climate draws attention to a number of non-linear changes to the cryosphere that are under way as the climate warms (Collins et al., 2019). Sea ice loss and melting of the Greenland Ice Sheet have sometimes been described as "tipping elements," vulnerable to irreversible decline beyond a particular climate threshold (e.g., Lenton et al., 2008; Ridley et al., 2010; King et al., 2020). Permafrost and mountain glaciers also have climatic thresholds for their long-term viability, while snow could be described as the definitive threshold feature of the climate system, existing below 0°C but in liquid phase above this value. Ecological and human systems have some degree of adaptive capacity and resilience to warming temperatures, but for snow and ice, the dependence on temperature is abrupt and non-negotiable.

The concept of irreversible climatic thresholds or tipping points is nonetheless uncertain in application to the global cryosphere. Snow and ice are present across a continuum of temporal and spatial scales, which smooths the cryospheric response to climate change. As a specific example, different sectors of the Greenland Ice Sheet have different climate sensitivities, so there can be large-scale decline of the ice sheet without a complete loss of ice. The survival of the central dome of the ice sheet in the last interglacial period is testament to this, as the core of the ice sheet proved resilient to temperatures 6 ± 2°C warmer than present, sustained for several millennia (NEEM Community Members, 2013; Yau et al., 2016). That is not to say that the Greenland Ice Sheet was insensitive to this warming; Yau et al. (2016) estimate that the ice sheet lost about 70% of its current volume, the equivalent of ∼5 m of global sea level, before the ice sheet fully re-established itself during the early stages of the last glaciation. Hence, the Greenland Ice Sheet was greatly impacted by the Eemian warming, but ice sheet retreat was neither inexorable nor irreversible.

Most elements of the cryosphere are immediately responsive to climate fluctuations, inclusive of changes that drive either the accumulation or the loss of snow and ice (e.g., Armour et al., 2011). The Greenland Ice Sheet is projected to reach a condition of negative surface mass balance this century (e.g., van den Broeke et al., 2016; Muntjewerf et al., 2020), which will accelerate ice sheet thinning and decline. However, the complete demise of the ice sheet requires many centuries to millennia (e.g., Huybrechts and De Wolde, 1999; Aschwanden et al., 2019), and could be stabilised or reversed under cooler climate conditions in future centuries. Mountain glacier melt, sea ice decline, and permafrost thaw are all reversible as well, although there may be a lag of years to decades after the climate first stabilises, due to the current state of disequilibrium. Hysteresis in the system can also make it more difficult to recover from the decline or loss of some elements of the cryosphere (e.g., Ridley et al., 2010; Levermann and Winkelmann, 2016).

While cryospheric changes are reversible in principle and tipping points are not well-defined, there is no question that snow and ice lack resilience to a continually warming climate. Most parts of the cryosphere have responded dramatically to recent climate change (IPCC, 2019), and will continue to do so if subject to ongoing warming. There are also many non-linearities and "local" thresholds associated with the loss of snow and ice, associated with feedbacks and regime changes in the cryosphere-climate system. This article outlines several such processes associated with glacier and ice sheet melt. Improved monitoring and modelling of these regime changes will increase confidence in projections of glacier and ice sheet response to climate change and the associated consequences for sea level rise, water resources, and mountain hazards.

FIELD SITES AND METHODS

Field Sites

Examples of regime shifts in glacier and ice sheet response to climate change presented in this article are assessed from recent literature, involving established but incompletely-understood processes. I also draw from field studies of glacier-climate processes carried out at Haig and Kwadacha Glaciers in the Canadian Rocky Mountains, Kaskawulsh Glacier in the St. Elias Mountains, and DYE-2, Greenland, on the southwestern flank of the Greenland Ice Sheet (**Figure 1**, **Table 1**).

Glaciological and meteorological studies at Haig and Kwadacha Glaciers in the Canadian Rocky Mountains include measurement and modelling of surface energy and mass balance, climate downscaling studies, glacier flow modelling, and glacio-hydrological process studies (Marshall, 2014; Ebrahimi and Marshall, 2015). Automatic weather stations (AWSs) were installed in the glacier forefield and in the upper ablation zone of each glacier for several years, providing insight into high-elevation off- and on-glacier meteorological conditions as well as information concerning the lapse rates (elevation dependence) of different meteorological fields. Both sites have mixed continental and maritime influences, with precipitation sources dominated by Pacific air masses that transport moisture to the Canadian Rocky Mountains.

Additional data discussed within this paper are drawn from field studies at Kaskawulsh Glacier, Yukon Territory, Canada and DYE-2, Greenland over the period 2016–2018. These studies focus on firn processes, based on shallow firn coring (up to 35 m depth), field measurements of meltwater infiltration and refreezing, and numerical modelling of the coupled decadal-scale evolution of firn temperature, hydrology, density, and ice content in response to climate change.

Firn cores and firn modelling on Kaskawulsh Glacier are described in Ochwat et al. (2021). Kaskawulsh is a large (1,095 km^2) glacier system within the St. Elias Mountains, and the field site for the firn cores is at an elevation of 2,640 m, in the upper accumulation area of the glacier. The site receives high amounts of precipitation delivered by Pacific storm tracks, ∼1.8 m w.e. yr^{-1}, and has mean annual and mean summer (JJA) temperatures of about −11 and −2°C, respectively (Ochwat et al., 2021). The ablation zone of Kaskawulsh Glacier is temperate (ice is at the pressure melting point), but it is unknown whether conditions are polythermal or temperate in the upper accumulation area. The extent to which meltwater is refrozen and retained on the upper glacier is also uncertain.

DYE-2, Greenland is a very different glaciological setting, in the upper percolation zone of the southwestern Greenland Ice Sheet. A GC-Net automatic weather station (AWS) has been operational at this site since 1997 (Steffen and Box, 2001), and this location has been the subject of several firn studies in recent years (Machguth et al., 2016; Heilig et al., 2018; MacFerrin et al., 2019; Samimi et al., 2020; Vandecrux et al., 2020b). DYE-2 is located near the Arctic circle, at an elevation with relatively cold, dry conditions. Mean annual accumulation is ∼0.36 m w.e. yr^{-1} (Mosley-Thompson et al., 2001), with mean annual and mean summer temperatures of about −18 and −5°C. The site is in

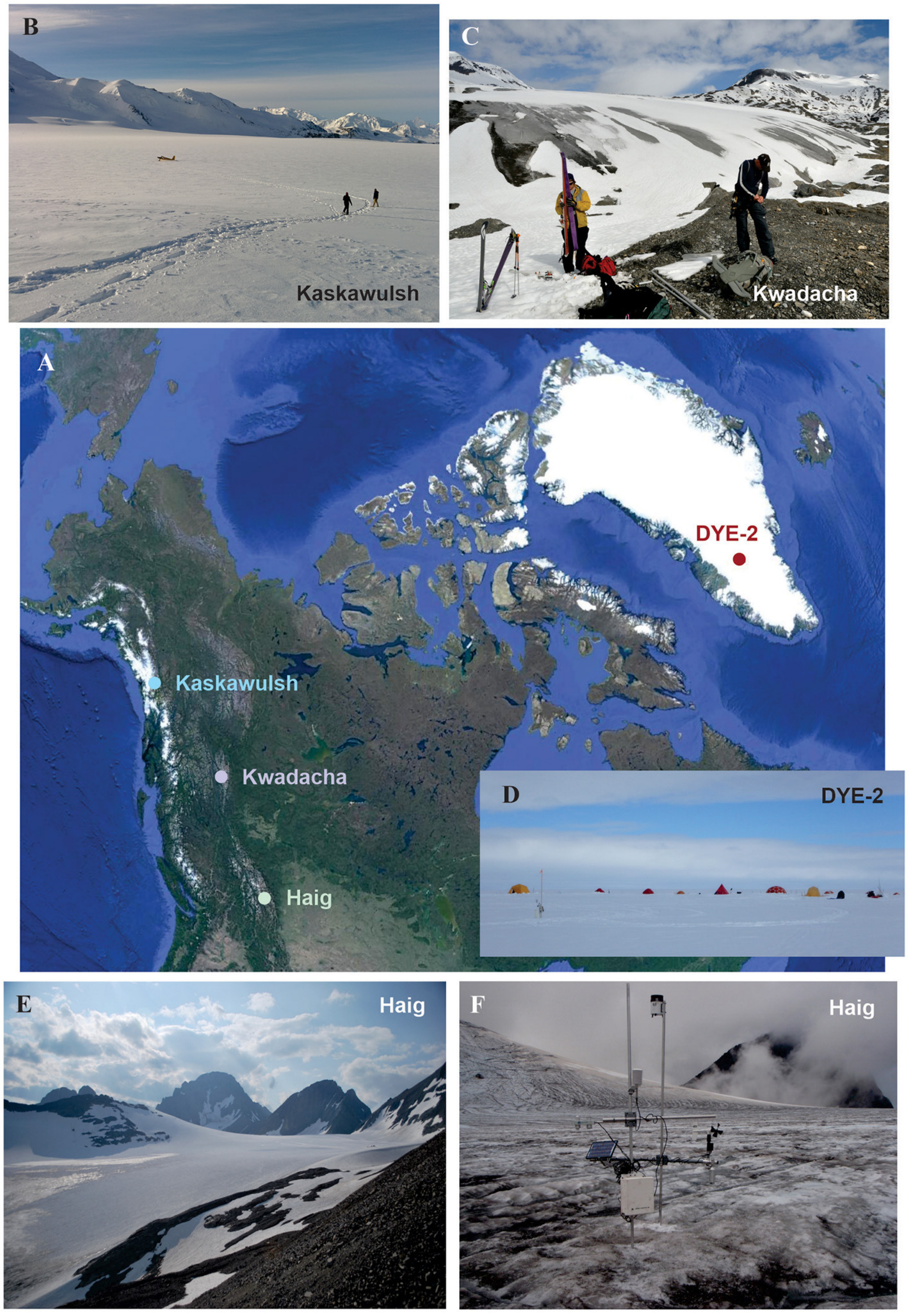

FIGURE 1 | (A) Glacier locations for data used in this study and photographs depicting the general environment of each setting: **(B)** Kaskawulsh Glacier, St. Elias Mountains, **(C)** Kwadacha Glacier, Canadian Rocky Mountains, **(D)** DYE-2, Greenland Ice Sheet, and **(E,F)** Haig Glacier, Canadian Rocky Mountains, with **(F)** depicting the automatic weather station on the upper glacier. Photograph **(C)** is courtesy of David Roberts, photograph **(E)** is by Chris Szymiec, and all others by the author.

the accumulation area of the ice sheet, with low summer melt totals and sufficient pore space and cold content to refreeze and retain 100% of meltwater (Machguth et al., 2016; Vandecrux et al., 2020b). Firn cores and modelling at this site augur its likely 21st-century transition from the accumulation area to the ablation area of the ice sheet, which would be accompanied by a transition from meltwater retention to runoff.

Methods

Meteorological, surface mass balance, snow/firn hydrology, and firn-core data collected from these field sites. These data are used to constrain models of surface energy balance, snow and ice melt, and the coupled evolution of snow/firn thermodynamics and hydrology at each site. Details on the model physics and parameterizations are presented in Ebrahimi and Marshall

TABLE 1 | Field sites drawn from in the analyses of glacier-climate regime changes.

Site	Latitude	Longitude	z (m)	References
Haig Glacier	50°43′N	115°18′W	2,665	Marshall, 2014
Kwadacha Glacier	57°50′N	124°57′W	2,005	Ebrahimi and Marshall, 2015
Kaskawulsh Glacier	60°47′N	139°38′W	2,640	Ochwat et al., 2021
DYE-2, Greenland	66°29′N	46°18′W	2,120	Samimi et al., 2020

For Haig and Kwadacha glaciers, the coordinates refer to locations where automatic weather stations were established on the glacier for multi-year periods. At Kaskawulsh and DYE-2 these refer to the locations of firn cores and firn hydrology process studies.

(2016), Samimi et al. (2020), and Ochwat et al. (2021), and are summarised here to support the results and analysis in this study.

AWS data include incoming and outgoing longwave radiation, $Q_L{}^{\downarrow}$ and $Q_L{}^{\uparrow}$, incoming and reflected shortwave radiation, $Q_S{}^{\downarrow}$ and $Q_S{}^{\uparrow}$, air temperature, relative humidity, wind speed, barometric pressure, and snow surface height. Variables were typically measured at 10-s intervals, with 30-min averages saved to Campbell Scientific dataloggers. In addition to these measurements, vertical arrays of thermistors, and time-domain reflectometer (TDR) probes were installed in the supraglacial snow firn during field experiments at Haig Glacier and DYE-2, Greenland, to study the subsurface thermal and hydrological evolution through the melt season (Samimi and Marshall, 2017; Samimi et al., 2020). TDR probes measure dielectric permittivity, which is a proxy for liquid water content in the snow and firn. These data were also recorded continuously and saved to Campbell Scientific dataloggers.

For reconstructions of the long-term evolution of these glacier systems we use hourly meteorological data from ERA5 surface-level climate reanalyses for the period 1950–2020 (Hersbach et al., 2020), based on the ERA5 grid cell over the point of interest. ERA5 has a resolution of 0.25° (∼28 km), so this is regionally and not locally representative. We therefore bias-adjust the ERA5 meteorological forcing using AWS data from each glacier. Where the AWS or ERA5 meteorological data are distributed over the glacier (horizontal distances of up to a few km), additional elevation adjustments are made using lapse rates for temperature and incoming longwave radiation, based on the elevation gradients measured on Haig and Kwadacha Glaciers: $dT/dz = -6°\text{C km}^{-1}$, $dQ_L{}^{\downarrow}/dz = -15\,\text{W m}^{-2}\text{ km}^{-1}$. Air pressure is adjusted following $dP/dz = -\rho_a g$, for gravitational acceleration g and air density $\rho_a = P/RT$. The gas law constant $R = 287\text{ J kg}^{-1}\text{ K}^{-1}$. In the absence of systematic observed variations, incoming shortwave radiation, wind speed, and relative humidity (RH) are assumed to be constant. Specific humidity is estimated from RH and from local air temperature and pressure.

Where upper-air analyses are included in this study, these are based on monthly mean ERA5 pressure-level data from 1950 to 2020. For the purpose of plotting the pressure-level data, altitude z_k is estimated from the pressure level, P_k, using ERA5 monthly temperature at each level, T_k. Temperature is used to calculate air density, $\rho_k = P_k/RT_k$, for the gas law constant R. The hydrostatic equation is then applied to estimate the elevation difference between pressure levels, $dz = -dP_k/\rho_k g$.

Meteorological and radiation data are used in the calculation of the surface energy balance,

$$Q_N = Q_S{}^{\downarrow} - Q_S{}^{\uparrow} + Q_L{}^{\downarrow} - Q_L{}^{\uparrow} + Q_H + Q_E + Q_C, \quad (1)$$

where Q_N is the net energy and Q_H, Q_E, and Q_C are the sensible, latent, and conductive heat fluxes, respectively, and energy fluxes are defined to be positive when they are directed toward the glacier surface. The surface energy balance model is coupled with a multi-layer subsurface model that simulates the snow, firn, and glacier temperature. Temperatures in the upper three layers are used in the calculation of the conductive heat flux, $Q_C = -k_t\, dT/dz$, for thermal conductivity k_t. The temperature of the surface layer is also used for the surface temperature and specific humidity in the sensible and latent heat flux calculations (Ebrahimi and Marshall, 2016).

When AWS data are available from the glacier, outgoing longwave, and shortwave radiation fluxes are directly available. When using climate reanalyses to drive Equation (1), these fluxes are parameterized. Outgoing longwave radiation follows Stefan-Boltzmann's equation, $Q_L{}^{\uparrow} = E_s \sigma T_s{}^4$, for surface emissivity $E_s = 0.98$, Stefan-Boltzmann's constant σ, and absolute temperature of the surface layer, T_s. Reflected shortwave radiation $Q_S{}^{\uparrow} = Q_S{}^{\downarrow}(1 - \alpha_s)$, for surface albedo α_s. The albedo is modelled to decline through the melt season as a function of cumulative positive degree days, PDD,

$$\alpha_s = \alpha_{s0} - k_\alpha \sum PDD, \quad (2)$$

where α_{s0} is the initial (spring) albedo and k_α is a decay constant used in the modelling, which is calibrated from observed albedo data at each study site. This parameterization is meant to represent observed reductions in snow albedo that occur through the melt season due to several effects, including increases in snow grain size, rounding of grains, liquid water content, and the progressive concentration of impurities in melting snow (Brock et al., 2000). Fresh-snow events can temporarily reset the surface albedo to the dry-snow value, α_{s0}, during the summer melt season. I treat this as a stochastic process, simulating random summer snowfall events as described in Marshall and Miller (2020). Once the fresh snow has melted, the surface albedo is restored to its pre-snow value.

When the surface layer is at 0°C and $Q_N > 0$, net energy in Equation (1) goes to melting, following

$$\dot{m} = Q_N/(\rho_w L_f), \quad (3)$$

where $\dot{m}$ is the melt rate (m s^{-1}), and L_f is the latent heat of fusion. If net energy is negative and the surface layer is 0°C, any liquid water that is present will refreeze, releasing latent heat, and the surface layer can then cool once all liquid water is refrozen. When surface layer temperatures are below the melting point, and surplus or deficit of energy drives warming or cooling, based on the energy balance solution within a one-dimensional model of subsurface temperature evolution:

$$\rho_b c_b \frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(-k_t \frac{\partial T}{\partial z}\right) + \varphi_t + \rho_w c_w q_w \frac{\partial T_w}{\partial z}. \qquad (4)$$

Here ρ_b, c_b, and k_t are the bulk density, specific heat capacity, and thermal conductivity of the snow or firn, ρ_w and c_w are the density and specific heat capacity of water, and q_w is the vertical rate of water percolation (m s^{-1}), with water temperature T_w. The right-hand terms in Equation (4) represent heat conduction, latent heat release from meltwater refreezing, and heat transport from advection of meltwater or rainwater, respectively. The thermal conductivity of the snow and firn is calculated after Calonne et al. (2019). The refreezing term in Equation (4) has units W m^{-3} and is calculated from

$$\varphi_t = \frac{\rho_w L_f \dot{r}}{z}, \qquad (5)$$

where $\dot{r}$ is the refreezing rate (m s^{-1}) and this heat is spread across the layer thickness, Δz.

The subsurface temperature model is coupled with a simple treatment of meltwater percolation. For a snow or firn layer with thickness Δz and the volume fraction of liquid water θ_w, the amount of water in the layer (m) is equal to $\theta_w \Delta z$. Conservation of mass in each subsurface layer gives the expression for local water balance,

$$\frac{d\theta_w}{dt} = -\nabla \cdot q_w - \frac{\dot{r}}{z}. \qquad (6)$$

Water that refreezes is assumed to be distributed over the layer Δz. Refreezing in Equation (6) is calculated within the subsurface thermal model, as a function of cold content in the layer. The water balance is forced at the upper boundary, where q_w is equal to the melt rate. Below this, meltwater fluxes are calculated from Darcy's law,

$$q_w = -k_h\left(\theta, \theta_w, d_g\right)\nabla\varphi, \qquad (7)$$

where $\nabla\phi$ is the hydraulic gradient and the hydraulic conductivity k_h is a function of the porosity, θ, liquid water content, θ_w, and snow/firn grain size d_g. The parameterization of hydraulic conductivity for unsaturated snow and firn follows that of Meyer and Hewitt (2017). Liquid water can also be retained within the pore space due to capillary forces, calculated after Coléou and Lesaffre (1998). Only excess water, beyond the irreducible water that is retained by capillary tension, is subject to vertical infiltration following Equation (7). Within this study, I assume purely vertical (gravitational) flow, such that the divergence and gradient terms in Equations (6) and (7) are replaced by vertical derivatives.

Ice layers in snow and firn can serve as impermeable barriers to meltwater infiltration, although this is not well-understood. Meltwater infiltration has been measured through continuous ice layers up to $\sim$0.12 m in thickness (Samimi et al., 2020). Infiltration through such layers may depend on fractures or discontinuities (Humphrey et al., 2021), which could develop in association with diurnal cycles of thermal expansion and contraction in near-surface environments. Meltwater could also potentially erode such layers. In contrast, ice slabs with thicknesses of $\sim$0.5 m or more may be impenetrable (Gascon et al., 2013; MacFerrin et al., 2019). In modelling presented here, I assume that ice layers <0.1 m thick are transmissive and ice layers more than 0.5 m thick act as impermeable barriers (k_h = 0). For ice layers with thickness t_i between 0.1 and 0.5 m, I prescribe a non-linear decrease in permeability by a factor $\kappa_i = 10^{-\gamma(ti-0.1)}$, with a reference value $\gamma = 10$. This reduces k_h by four orders of magnitude as ice-layer thickness increases from 0.1 to 0.5 m. Sensitivity experiments presented in section From Meltwater Retention to Runoff explore the effects of permeable vs. impermeable ice layers.

REGIME SHIFTS IN GLACIERS AND ICE SHEET RESPONSE TO CLIMATE CHANGE

From Polythermal to Temperate Firn and Ice

The thermal regime of a glacier mass is a fundamental distinction in glaciology. Outside of the tropics, the upper several metres of ice, firn, and snow on a glacier typically experiences seasonal freezing through heat loss to the atmosphere. Through thermal diffusion, this results in a winter temperature wave that propagates to depths of up to $\sim$10 m (Cuffey and Paterson, 2010). Below this, *temperate* glaciers are at the pressure-melting point throughout, while *polythermal* glaciers have a range of temperatures (Blatter, 1987; Blatter and Hutter, 1991). Ice temperatures in polythermal glaciers are generally warmer near the bed, due to the combined effects of geothermal heat fluxes and heat dissipation from creep deformation of the ice, which is concentrated in the lower glacier (Cuffey and Paterson, 2010). Heat can also be generated at the bed from frictional heat dissipation, where the glacier is sliding over the substrate.

In some cases, polythermal ice is warm-based (at the pressure-melting point), enabling the presence of liquid water in thermal equilibrium and supporting basal flow through either sliding or subglacial sediment deformation (Clarke, 1987). In other cases, polythermal ice masses can be frozen to the bed, a condition known as cold-based ice. Freezing conditions can be maintained if heat conducted upwards through the ice exceeds inputs of energy from geothermal heat flux, creep deformation, and frictional heat dissipation. Polar glaciers, icefields, and ice sheets are typically polythermal and have a mixture of warm- and cold-based ice. Basal ice temperatures do not follow a simple relationship with mean annual air temperature, so it is difficult to predict whether high-latitude or high-altitude glaciers and ice sheets will be warm- or cold-based. For instance, much of the plateau of the East Antarctic Ice Sheet, the coldest region in the world, is underlain by warm ice (e.g., Bell et al., 2007), because the ice is thick enough ($\sim$4 km) to insulate the base from cold surface temperatures. In contrast, the interior of the Greenland Ice Sheet is frozen to the bed (Huybrechts, 1996; Marshall, 2005) because it is thinner ($\sim$3 km) and snow accumulation rates are higher than in Antarctica. This results in effective downward advection of cold surface ice in Greenland, increasing the temperature gradient (hence, conductive energy loss) in the lower part of the ice sheet. On the flanks and in the marginal areas of Greenland, most of the ice sheet is believed to be warm-based, due to the combined influences of warmer air temperatures and strain heating in areas of active vertical shear deformation. Latent heat release from refreezing meltwater in the surface snow and firn also plays a significant thermodynamic role in parts of the Greenland Ice Sheet, as discussed in this study.

The thermal regime of glaciers is important for several reasons. Warm-based ice can flow more quickly via basal flow or ice surging (e.g., Clarke, 1987; Sevestre et al., 2015), and also enables a range of subglacial hydrological processes that differ from cold-based ice (e.g., exchanges with the groundwater system; chemical weathering, erosion, and mobilisation of sediments in well-developed drainage systems; subglacial water storage). Warmer ice has a lower effective viscosity (Glen, 1955), with variations by a factor of ∼1,000 from temperatures of −50 to 0°C (Marshall, 2005); hence, deformational velocities in temperate ice are much higher (e.g., Phillips et al., 2010; Bell et al., 2014). Meltwater also cannot refreeze in temperate ice and firn, limiting the extent of meltwater retention through refreezing (Pfeffer et al., 1991), although meltwater can be retained as liquid water in temperate firn where there is adequate infiltration and pore space (e.g., Koenig et al., 2014; Christianson et al., 2015). Processes of meltwater refreezing and retention are important to glacier and ice sheet mass balance (section From Meltwater Retention to Runoff).

The transition from polythermal to temperate ice is expected in some glacier environments as a consequence of climate warming (e.g., Hoelzle et al., 2011). Sub-polar glaciers are good candidates to experience this transition in the coming decades, as these systems often lie close to the climatic threshold between polythermal and temperate conditions. Firn temperature observations on Lomonosovfonna, Svalbard (Marchenko et al., 2021) indicate that temperatures at ∼12-m depth may have shifted from sub-freezing to temperate conditions over the past two decades. Such a transition may also be close at high elevations in the European Alps, where warming of cold firn has been measured over the last two decades (e.g., Hoelzle et al., 2011; Mattea et al., 2021; Colle Gnifetti on Monte Rosa). Climate change trends are strong at high latitudes (e.g., Zhang et al., 2019) and high elevations (Pepin et al., 2015; Williamson S. et al., 2020), which is likely to accelerate regime changes from polythermal to temperate ice.

Ochwat et al. (2021) report on a possible transition from polythermal to temperate conditions that was discovered during ice-coring work in May 2018 on the upper Kaskawulsh Glacier in the St. Elias Mountains. The drilling team was surprised to encounter temperate firn conditions with liquid water (i.e., a firn aquifer) at ∼34 m depth. Based on density measurements in the firn core, this is near the base of the firn; the aquifer appears to be a few metres thick, representing a water table that is perched on the glacier ice. It is unclear whether or not the firn aquifer is new. Previous radar studies and ice-coring efforts in the region (Zdanowicz et al., 2014) provide no prior indication of the presence of perennial firn aquifers. It is possible that temperate conditions have existed for many decades at this site, but Ochwat et al. (2021) argue that the presence of deep, temperate, and water-saturated firn is a new development on Kaskawulsh Glacier.

Firn modelling at the drill site supports this hypothesis. Extending the results reported by Ochwat et al. (2021), **Figure 2** plots simulated firn temperature evolution on upper Kaskawulsh Glacier for the period 1950–2020, using the coupled thermodynamic and hydrological model described in section Methods for the upper 35 m of firn. ERA5 meteorological fields are bias-adjusted using data from a nearby automatic weather station (Ochwat et al., 2021), and initial firn temperature and stratigraphy are based on a 30-year spin-up using the ERA5 climatology from 1950 to 1979. Based on these initial conditions, the climate forcing was restarted at 1950 for the 71-year simulation plotted in **Figure 2**.

Mean annual air temperature at the site over the period 1950–2020 was −11.2°C, increasing to an average of −10.3°C for the last decade (2010–2020), while mean summer (JJA) temperatures shifted from a long-term mean of −2.8 to −1.8°C for the last decade (**Figure 2A**). This decadal-scale warming was accompanied by increases in atmospheric humidity and incoming longwave radiation, as reported by Williamson S. et al. (2020), with overall increases in net energy driving an increase in surface melting at a rate of +0.03 m w.e. per decade (**Figure 2B**). The resulting intensification of meltwater infiltration and refreezing drove decadal-scale firn warming (**Figures 2C–E**), culminating in a transition from polythermal to temperate conditions in 2013–2014. Meltwater refreezing since this time provides sufficient latent heat release to maintain temperate firn, despite a mean annual air temperature of about −10°C. The climate conditions and firn processes governing this abrupt regime change are discussed in section Processes Governing Nonlinear Glacier and Ice Sheet Response to Climate Change.

This is a marked regime change for the glacier, as the glacier ice that forms from the firn will now develop into temperate ice. Glacier ice advecting downwards and downslope from the accumulation area will be more deformable than ice from earlier decades, when the firn column was cold. The importance of englacial and subglacial temperatures for ice dynamics is discussed in section Englacial and Subglacial Warming.

From Meltwater Retention to Runoff

Prior to 2013 at Kaskawulsh Glacier, all surface melt refroze within the firn (100% meltwater retention; **Figure 2B**). In summers with deep meltwater infiltration, liquid meltwater that is stored in the firn may not refreeze until the subsequent calendar year, as this requires penetration of the following winter's cold wave into the firn. Melting, refreezing, and runoff are calculated for the calendar year, with runoff calculated from melting minus refreezing. The lag in refreezing can give years with small amounts of positive runoff followed by years with "negative" runoff in **Figure 2B**; this is not a violation of conservation of mass, but an artefact of calculating this for calendar years.

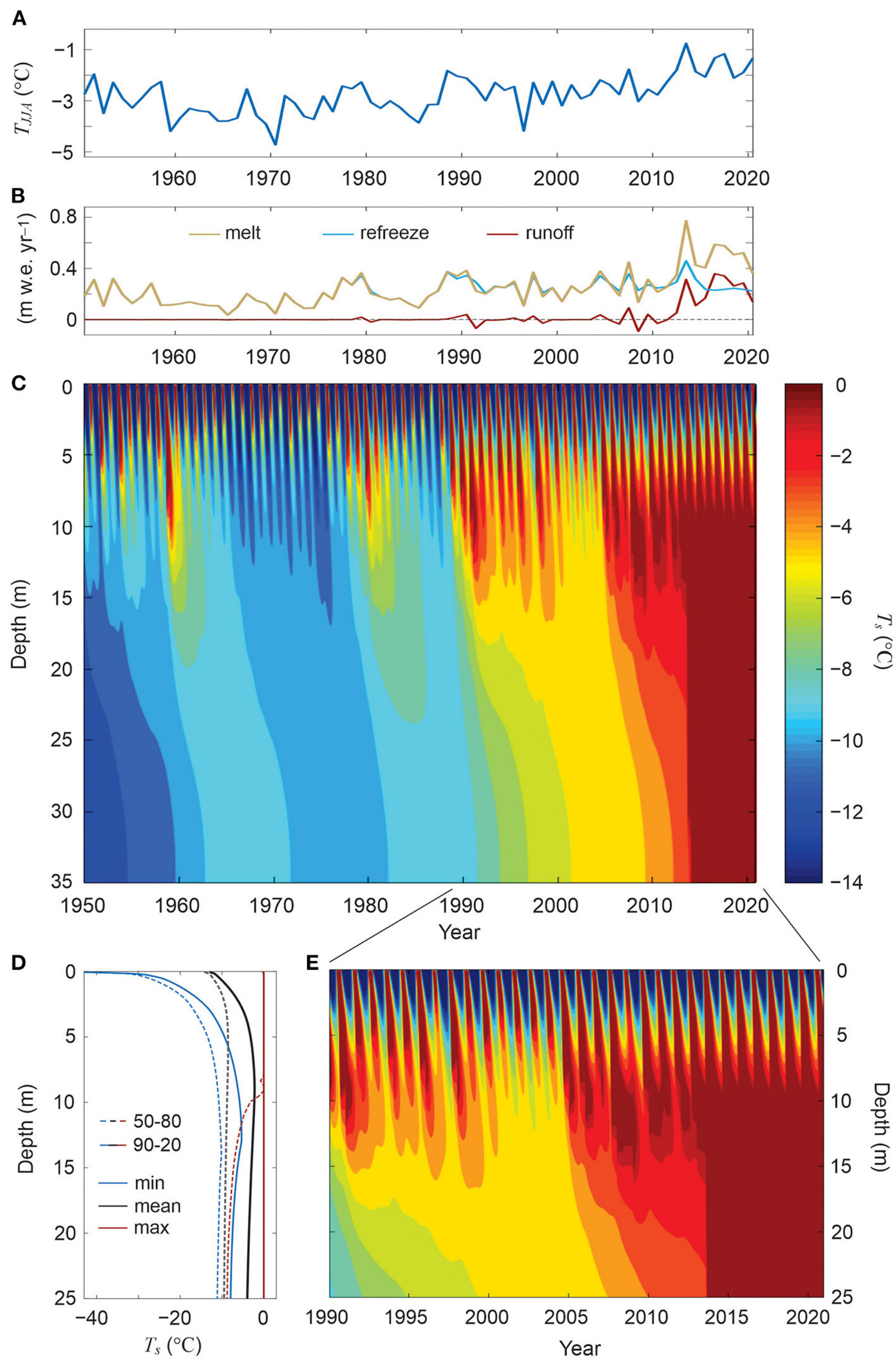

FIGURE 2 | Modelled temperate firn development at Kaskawulsh Glacier, Yukon Territory over the period 1950–2020. **(A)** Mean summer air temperature in the ERA5 climate forcing. **(B)** Annual melt, refreezing, and runoff. Runoff is calculated from the annual melt minus refreezing. **(C)** Firn temperatures to 35 m depth. **(D)** Firn temperatures in the upper 25 m, 1990–2020 **(E)** Firn temperature envelope in the upper 25 m for the periods 1950–1980 (dashed) and 1990–2020 (solid lines). "Negative" runoff in **(B)** occurs when meltwater drainage to the deep firn from the prior summer refreezes in the subsequent calendar year.

Following the intensive melt season of 2013, which triggered the transition to temperate firn on Kaskawulsh Glacier, meltwater has begun to run off from this site (**Figure 2B**). Some summer meltwater still refreezes within the winter cold layer, which extends to about 8 m depth (**Figures 2D,E**), but for an average melt rate of 0.47 m w.e. yr^{-1} from 2010 to 2020, 0.18 m w.e. yr^{-1} (39%) is modelled to have drained to the deep firn. About half of this meltwater was retained within the firn aquifer (Ochwat

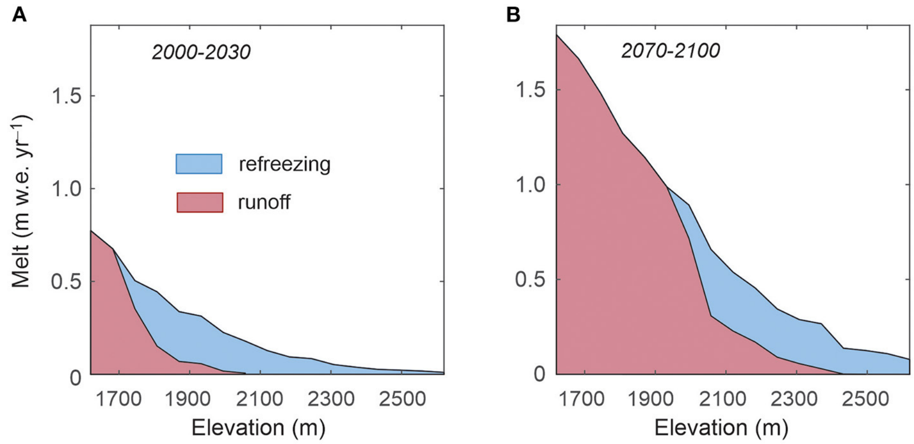

FIGURE 3 | Mean annual surface melt rates as a function of elevation for a transect at 66.5°N in southwest Greenland for **(A)** 2000–2030 and **(B)** 2070–2100. The red shading indicates the surface runoff and the blue shading indicates the meltwater retained by refreezing.

et al., 2021), while the remainder is expected to have drained laterally, either as Darcian flow in the firn aquifer or at the firn-ice interface, where the glacier ice presents an impermeable barrier. This transition therefore impacts meltwater retention and surface mass balance on the upper Kaskawulsh Glacier.

This is an example of a wide-spread trend in the world's glaciers, ice caps, and ice sheets. Meltwater retention capacity is declining in response to loss of firn (e.g., Pelto, 2006) and increases in firn temperature, density, and ice content, all of which reduce the available cold content and pore space that is available for meltwater refreezing (e.g., van Angelen et al., 2013; Vandecrux et al., 2020a). Pfeffer et al. (1991) and Harper et al. (2012) describe the importance of meltwater refreezing and retention in the Greenland Ice Sheet. The firn layer acts as a buffer to sea level rise, as a large amount of meltwater stays within the ice sheet rather than running off to the ocean. Temperate snow, firn, and ice have limited capacity for meltwater retention, since liquid water cannot refreeze. Meltwater can be stored in the pore space of temperate snow and firn, but this generally drains effectively through gravitational percolation or Darcian flow in firn aquifers (e.g., Christianson et al., 2015), and water will run off once the available pore space is filled. Increases in meltwater runoff also result from reduced permeability in firn, through densification, and the development of thick near-surface ice layers (de la Peña et al., 2015; MacFerrin et al., 2019).

Noël et al. (2017) describe this transition from meltwater retention to runoff as a tipping point for glacier mass balance—from positive to negative—in glaciers and ice caps peripheral to the Greenland Ice Sheet. A similar mass balance regime change may have occurred in Svalbard, where a shift from meltwater retention to runoff has caused a large area of Svalbard's ice caps to transition from net accumulation to net ablation (Noël et al., 2020). Reductions in capacity for meltwater retention in firn have also been documented on the Barnes and Devon Ice Caps in Arctic Canada (Zdanowicz et al., 2012; Bezeau et al., 2013). This transition may also be occurring in the lower percolation zone of the Greenland Ice Sheet (Charalampidis et al., 2015; de la Peña et al., 2015). High melt rates can produce thick near-surface ice slabs which limit meltwater infiltration and refreezing (Machguth et al., 2016; MacFerrin et al., 2019). The percolation zone in Greenland is expanding to higher elevations of the ice sheet, but there is still sufficient cold content to support meltwater refreezing and retention in the vast interior regions of Greenland (Vandecrux et al., 2020a), as long as the permeability of the near-surface firn continues to permit meltwater infiltration. There is concern that strong episodic melt seasons, as observed in 2012 and 2019, may create ice layers that limit infiltration, and model projections suggest that Greenland could experience significant losses of firn meltwater retention capacity by the second half of this century (van Angelen et al., 2013).

Figure 3 plots the modelled increase in melting and meltwater runoff as a function of elevation in a simple future warming scenario on the southwestern flank of the Greenland Ice Sheet. This scenario is based on ERA5 climatology for the period 2000–2020 (Hersbach et al., 2020), followed by a linear warming rate of 0.5°C per decade from 2020 to 2100, superimposed on the climatology of the base period 2000 to 2020. This is a simplistic scenario, intended to examine the sensitivity of melting and meltwater retention processes to a hypothetical warming of 4°C in Greenland by the end of the century. The climate forcing drives a surface energy balance and firn model that have been calibrated for DYE-2 in southwest Greenland (Samimi et al., 2020), and the elevation range in **Figure 3** is representative of the present-day percolation zone in southwest Greenland.

For the period 2000–2030, the average melt rate from 1660 to 2640 m elevation in **Figure 3A** is 0.23 m w.e. yr^{-1}. This increases to 0.72 m w.e. yr^{-1} for the period 2070–2100 (**Figure 3B**). The shading in **Figure 3** indicates the fraction of meltwater runoff relative to the total meltwater production. Under a 4°C warming by end of the century, there is a transition from meltwater retention to runoff in the elevation range from ∼1,800 to 2,100 m. The equilibrium line and percolation zone climb to higher

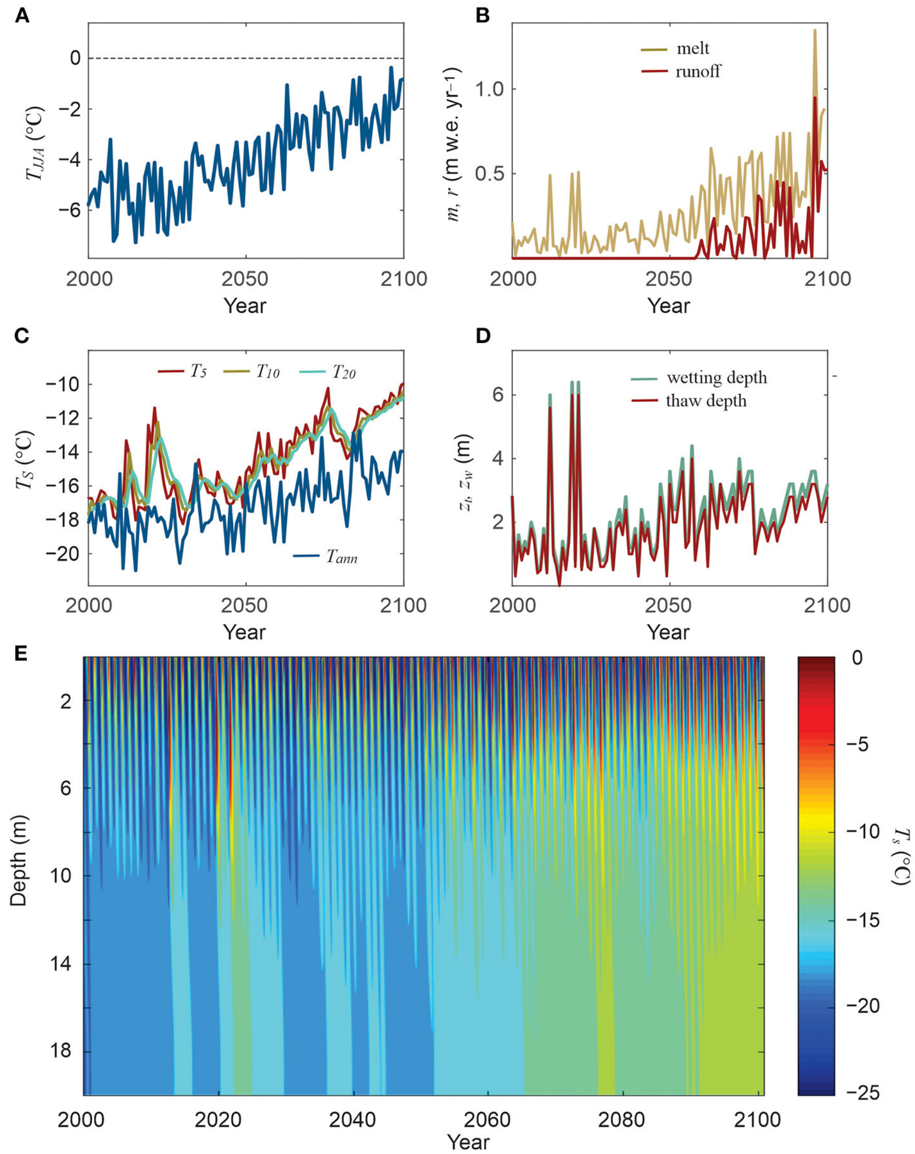

FIGURE 4 | Modelled firn evolution at DYE-2, Greenland, 2000–2100, using ERA5 climate forcing from 2020 to 2020 and with a warming trend of 0.05°C yr^{-1} for 2020–2100. **(A)** Mean summer air temperature. **(B)** Annual surface melt and runoff. **(C)** Mean annual snow/firn temperatures at the surface and at 5-, 10-, and 20-m depth. **(D)** Maximum annual thaw and meltwater infiltration depths. **(E)** Daily mean temperatures in the 20-m firn column from 2000 to 2100.

elevations, and under the 4°C warming there is no longer a dry snow zone in southern Greenland. Above ∼2,440 m, however, almost 100% of meltwater is still retained in 2,100. For the total elevation range represented in **Figure 3**, average runoff for the period 2000–2030 equals 0.08 m w.e. yr^{-1} (33% of total melt). This increases to 0.45 m w.e. yr^{-1} (62% of total melt) for 2070–2100 (**Figure 3B**). Melt rates in the present-day percolation zone of southwestern Greenland increase by a factor of ∼3 under this warming scenario, but runoff increases by a factor of ∼6.

This is a non-linear but continual change for Greenland Ice Sheet mass balance and meltwater runoff, rather than a threshold process. However, a given location on the ice sheet may experience a regime change such as the transition from the accumulation area to the ablation area of the ice sheet, or from meltwater retention to runoff (Charalampidis et al., 2015), similar to what was observed on Kaskawulsh Glacier in **Figure 2**. The nature of this regime change depends on the climate and firn conditions. As an example, **Figure 4** plots the modelled firn temperature evolution at DYE-2, Greenland from 2000 to 2100 under the hypothetical warming scenario discussed in **Figure 3**. DYE-2 is situated at 2120 m elevation on the southwestern flank of the ice sheet. It is presently in the upper percolation zone, with moderate amounts of annual melt (∼0.15 m w.e. yr^{-1}), 100% of meltwater retained as refrozen

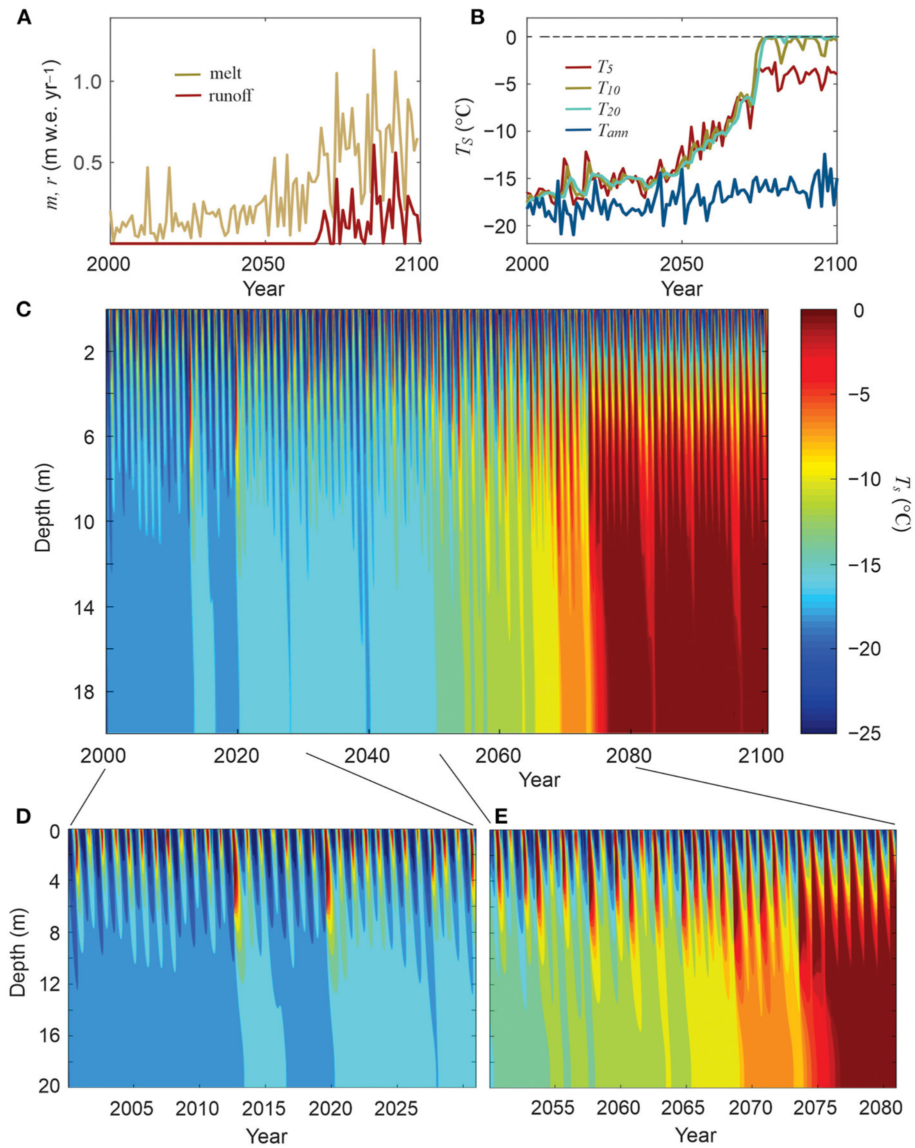

FIGURE 5 | Modelled firn evolution at DYE-2, Greenland as per **Figure 4** but prescribing a 50% increase in precipitation at DYE-2. **(A)** Modelled melt and runoff. **(B)** Mean annual air temperature and firn temperatures at 5-, 10-, and 20-m depth. **(C–E)** Daily mean temperatures in the 20-m firn column from 2000 to 2100, **(D)** 2000 to 2030, and **(E)** 2050 to 2080, showing the transition to deep temperate firn in 2077.

ice, a mean annual temperature of −17.4°C, and a 10-m firn temperature of about −15.5°C (Samimi et al., 2020; Vandecrux et al., 2020b). Under the imposed gradual warming scenario, the equilibrium line elevation climbs to ∼2,050 m by 2100 and DYE-2 remains within the accumulation area of the ice sheet, with mean annual and summer (JJA) temperatures of −14.4 and −2.3°C for the period 2070–2100 (**Figures 4A,C**). Firn at DYE-2 is projected to remain polythermal through the century (**Figure 4E**), with 10-m temperatures of about −10°C by 2100 (**Figure 4C**).

This projected firn evolution contrasts with that at Kaskawulsh Glacier for several reasons. In part, DYE-2 is colder, with less summer melt. This limits the latent heat release from meltwater refreezing, helping DYE-2 to remain polythermal. More importantly, however, deep meltwater infiltration events become less frequent in the second half of the century (**Figure 5D**), despite increasing melt rates (**Figure 5B**). Warming causes increased meltwater production in the 2050s, but much of this meltwater refreezes near the surface, driving the formation of dense firn with thick ice layers which impede meltwater infiltration. This triggers the onset of meltwater runoff from this site (**Figure 5B**). Hence, the site evolves from a location with 100% meltwater retention from 2000 to 2030 to an average meltwater retention of 58% from 2070 to 2100. Average net

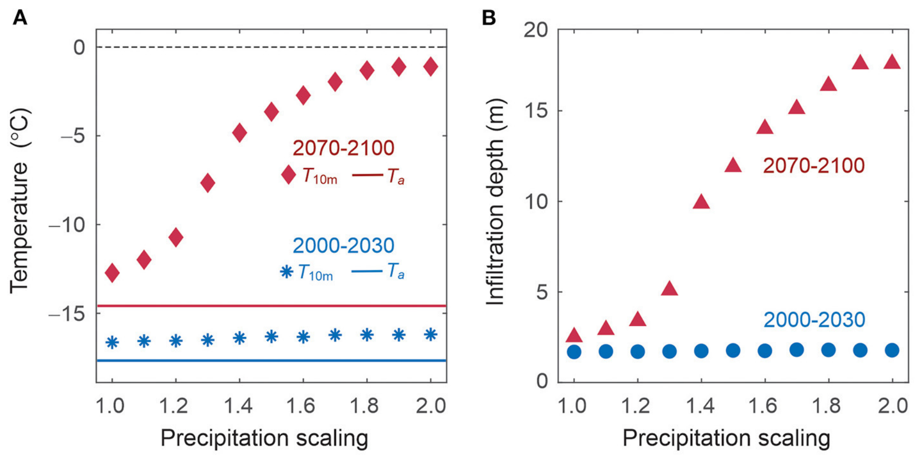

FIGURE 6 | Effects of snow accumulation rate on modelled 21st-century **(A)** 10-m firn temperature and **(B)** meltwater infiltration depth at DYE-2, Greenland. Blue symbols are average values for the period 2000–2030 and red symbols are for 2070–2100. Lines in **(A)** indicate the mean annual air temperature for each period. Temperatures and infiltration depths are plotted for precipitation rates up to twice the present-day value at DYE-2, giving winter mass balances $b_w \in [1, 2]$.

mass balance decreases from 0.38 to 0.16 m w.e. yr^{-1} over these two periods.

Limitations to meltwater infiltration are partly associated with the lack of near-surface cold content and pore space. As discussed by Kuipers Munneke et al. (2014), deep meltwater infiltration can be supported by high rates of snow accumulation, which increase the available pore space for meltwater refreezing and help to insulate the underlying firn, where liquid meltwater may be present from the previous melt season. As an illustration of this at DYE-2, snow accumulation rates (winter mass balance, b_w) were increased by 10% intervals from 100 to 200% of the present-day rate. The prescribed values of b_w range from an average of 0.37–0.74 m w.e. yr^{-1} for the reference period 2000–2030. Combined with the prescribed increase in mean annual precipitation of +5%°C^{-1} in the future warming scenario, the corresponding range in b_w for 2070–2100 is 0.44–0.87 m w.e. yr^{-1}.

The modelled 21st-century firn evolution is strongly dependent on accumulation rates (**Figures 5**, **6**). Present-day firn conditions do not vary much in these numerical experiments, even for a doubling of accumulation, but a ~40% increase in accumulation is sufficient to give a dramatically different outcome by end of century. **Figure 5** plots modelled firn temperature from 2000 to 2100 for a 50% increase in accumulation at DYE-2. All other climate and parameter settings in this experiment are identical to the baseline model shown in **Figure 4**. With the additional accumulation and meltwater refreezing capacity, firn at DYE-2 undergoes an abrupt transition from polythermal to temperate conditions in the late 2070s (**Figures 5B,C**). The transition is similar to the one observed on Kaskawulsh Glacier (**Figure 2**). Temperate firn from 10- to 20-m depth first develops in 2077, coinciding with meltwater infiltration through the full firn column. Modelled firn warming at 10-m depth is 1.2°C per decade, 3.4 times the rate of atmospheric warming (**Figure 5B**), although the firn warming is abrupt rather than linear. The winter cold wave at this site reaches a depth of about 8 m each year (**Figure 5C**), providing some meltwater refreezing capacity while this site remains in the accumulation area of the ice sheet. Similar to the baseline model in **Figure 4**, meltwater runoff from the site begins in the 2060s, associated with low-permeability near-surface ice layers.

Figure 6 plots summary diagnostics for the precipitation sensitivity experiments, showing firn temperatures and meltwater filtrations depths for the reference period, 2000–2030 (blue symbols) and the end of the century, 2070–2100 (red symbols). The onset of deep meltwater corresponds with precipitation increases of ~40% or more, attended by deep firn warming. For 2000–2030, 10-m firn temperatures are 1.3°C warmer than the mean annual air temperature, on average. This differs only slight for 2070–2100 with the present-day accumulation rate, but for b_w values from 150 to 200% of present-day values, 10-m firn temperature is 11–14°C warmer than the mean annual air temperature. The transition to temperate conditions in the deep firn is difficult to reverse though (see section Processes Governing Nonlinear Glacier and Ice Sheet Response to Climate Change), such that this regime change leads to long-term reductions in firn meltwater retention capacity.

Meltwater Refreezing on Mountain Glaciers

Meltwater retention through refreezing is not a significant process on most mountain glaciers, due to temperate conditions, but meltwater refreezing still plays an important and more subtle effect in reducing the net energy that is available for melt. Meltwater that is stored in the pore space of near-surface snow/firn or in melt ponds on the glacier surface commonly refreezes overnight or during cold-weather episodes. This is a familiar process to mountaineers, who travel quickly and lightly over the refrozen snow and ice crust in the early morning hours. Energy is used to warm this crust to the melting point the following morning, and some of this energy is then used to "remelt" the refrozen meltwater. A significant amount of melt energy can be consumed by this recycled meltwater over several months of diurnal freeze-thaw cycles.

Samimi and Marshall (2017) estimate this process to reduce meltwater runoff by about 10% at Haig Glacier in the Canadian Rockies. Although the glacier and the supraglacial snowpack are temperate, the thin refreezing layer at the glacier surface consumes energy for the diurnal phase changes. This is consistent with the observed lag of meltwater production on the glacier surface and in the proglacial stream on a typical summer day. Diurnal increases in supraglacial streamflow commonly don't set in until late morning or early afternoon. As the climate warms and the glaciers experience less frequent overnight refreezing, diurnal freeze-thaw cycles will diminish in importance. Glaciers that are losing their firn will also see reductions in meltwater storage capacity. Collectively, these processes will result in "flashier" hydrological systems with reduced meltwater residence times.

Englacial and Subglacial Warming

Meltwater that infiltrates beyond the firn layer can also refreeze in cold glacier ice, with latent heat release directly warming up englacial or basal ice (Jarvis and Clarke, 1974). In Greenland, water that penetrates to depth through crevasses, moulins, and proglacial lake drainage (Das et al., 2008; Catania and Neumann, 2010) has been proposed as a mechanism to soften ice and increase flow rates through a process called cryo-hydrologic warming (Phillips et al., 2010). Latent heat release produces warmer ice, which has a lower effective viscosity, and liquid water content can further soften temperate ice (Cuffey and Paterson, 2010). Cryo-hydrologic warming and temperate ice therefore support greater rates of ice deformation, resulting in glacier thinning. Higher flow rates also increase the advection of ice to low elevations or the ocean, where there is surplus energy for ablation. This creates a more negative glacier mass balance and increases meltwater runoff.

Melting conditions at the glacier or ice sheet bed have the potential to induce higher rates of flow through sliding or sediment deformation, where sufficient meltwater is available to support basal flow through these processes (Iken and Bindschadler, 1986; Clarke, 1987). The transition from cold- to warm-based ice therefore represents a regime change that can fundamentally alter glacier dynamics. Increased ice fluxes through basal flow have a similar impact to low-viscosity ice: thinner ice masses, more negative mass balance, and a shorter response time to climate forcing. A transition from cold- to warm-based ice has been proposed as the trigger for deglaciation in the context of the 100-kyr glacial cycle (Marshall and Clark, 2002). The effects of basal flow have the potential to be dramatic, as observed in glacier surge cycles, since basal flow rates can greatly exceed those from internal ice deformation (e.g., Clarke, 1987; Sevestre et al., 2015).

Surging glaciers and ice streams provide clear examples of thermally-enabled, meltwater-lubricated fast flow in glaciers and ice sheets, but it is difficult to assess whether anthropogenic climate forcing can trigger changes in flow regimes. Subglacial thermal and hydrological conditions would need to evolve over large areas of the bed on annual to decadal timescales. The diffusive and advective timescales for atmospheric temperature changes to reach the subglacial environment are long: centuries for mountain glacier and millennia for ice sheets. Surface meltwater that reaches the bed has the potential to incite rapid change, through both latent heat release and impacts on subglacial hydrology (e.g., Zwally et al., 2002). Subglacial water that does not drain effectively can become pressurised, reducing basal friction (ice-bed coupling), and enabling speedups (Parizek and Alley, 2004; Colgan et al., 2012; Shannon et al., 2013).

These processes are not fully understood. The fate of meltwater that drains through crevasses, moulins, and proglacial lakes in Greenland is uncertain; some may refreeze near the surface and some may drain effectively from the ice sheet, through either englacial or subglacial pathways. To have a major effect on flow rates or on viscous softening, this water needs to penetrate to depth and remain within the ice sheet system, where it can lubricate the ice sheet bed or where refreezing can effectively warm and soften the deep ice. Das et al. (2008) and Catania and Neumann (2010) demonstrate that supraglacial water in Greenland can effectively reach the bed, so these processes have the potential to impact the ice sheet as the ablation area and wet-snow zones expand to higher elevations in Greenland in future decades. Previously frozen or dry parts of the ice sheet bed may become newly exposed to meltwater, with the potential to increase flow rates (Parizek and Alley, 2004; Shannon et al., 2013). On the other hand, efficient subglacial hydrological drainage systems typically develop in settings where large amounts of meltwater reach the bed. This effectively evacuates subglacial water rather than supporting the distributed, high-pressure water systems that support fast flow (e.g., Schoof, 2010; Sundal et al., 2011).

The net impact of climate warming and increasing amounts of meltwater on glaciers and ice sheet dynamics is uncertain. Based on ice sheet modelling, Shannon et al. (2013) conclude that the impacts of changing hydrology on basal flow are likely to be negligible in Greenland this century, in comparison with the direct impacts of climate change on surface mass balance and sea-level rise. Davison et al. (2019) reach a similar conclusion based on a review of recent literature analysing the relation between supraglacial water drainage and seasonal ice acceleration in Greenland. Models of englacial drainage and subglacial hydrology are becoming increasingly sophisticated (Flowers, 2015; Banwell et al., 2016; Hoffman et al., 2016), but challenges of scale, complexity, and the dearth of direct observations in subglacial environments limit understanding and modelling of ice-sheet scale hydrological and basal flow processes. For these reasons, ice sheet dynamical response to changing ice sheet hydrology remains an open question, and the possibility of widespread impacts of increasing meltwater supply to the bed cannot be ruled out.

Thawing at the Ice-Rock Interface

The transition from frozen to thawed conditions at the glacier bed is mirrored by thawing of the mountains themselves: reduced mountain permafrost as temperatures rise and active layers deepen. This interacts with glacier melting, as newly-exposed rock surfaces and circulating meltwater act collectively to increase mountain hazards associated with slope instabilities (Hock et al., 2019). The transition from frozen to thawed conditions can trigger rock and ice detachment processes in high-mountain regions (Gruber and Haeberli, 2007; Gilbert et al., 2015; Hock et al., 2019), through a combination of factors including decreased structural integrity (loss of ice bonding), lubrication from liquid water, and mechanical and thermal stresses from freeze-thaw cycles (Fischer et al., 2012; Haeberli et al., 2017). These processes have been linked with increasing frequency of rockfall and rock avalanches (e.g., Allen et al., 2011; Ravanel and Deline, 2011; Fischer et al., 2012; Allen and Huggel, 2013; Hock et al., 2019), and may play a role in glacier detachments (Fischer et al., 2013; Faillettaz et al., 2015; Gilbert et al., 2015). High-mountain slope failures are often associated with a mixture of ice, rock, and water. Glacier retreat can also directly influence slope stability (e.g., Allen et al., 2011; Hock et al., 2019), as fresh walls and debris become exposed and buttressing of steep slopes is reduced.

From White to Grey

The collection *Darkening Peaks: Glacier Retreat, Science, and Society* (Orlove et al., 2008) provides a compelling image of the transition from white to grey that is under way in the world's mountain regions. The demise of glaciers is changing the fundamental character of mountain landscapes, with direct impacts on regional weather, climate, hydrology, ecosystems, and tourism. Regional surface albedo also declines due to the loss of snow and ice, which may be contributing to elevation-dependent warming in some of the world's mountain regions (Pepin et al., 2015).

Changes in surface albedo in mountain regions can be subtle, and are less "binary" than the shift observed in polar regions when sea ice is replaced with open water. Highly-reflective seasonal snow still covers mountains for much of the year at high elevations, but there is a strong decline in albedo when seasonal snow on glaciers melts away to expose the underlying glacier ice. This transition is important to the surface energy balance and melt, and a shorter snow season contributes to regional reductions in albedo. Without snow cover, most mountain glaciers can already be described as "grey." Glacier ice albedo varies from ∼0.1 to ∼0.6 (Bøggild et al., 2010; Cuffey and Paterson, 2010) but is commonly at the low end of this range on mountain glaciers. Low albedo values are associated with high concentrations of particulate matter on the ice, which can accumulate over many melt seasons. The photograph in **Figure 1F** illustrates this well, corresponding to an ice albedo of about 0.2.

There are numerous accounts of glaciers darkening in recent years due to increasing concentrations of particulate matter (e.g., Dumont et al., 2014; Ming et al., 2015; Tedesco et al., 2016; Stibal et al., 2017; Williamson et al., 2019). Impurities on glaciers have a variety of sources, some of which are linked to climate change. Mineral dust and aerosols make up the greatest mass fraction of particulate matter in many mountain glacier settings, associated with both local and long-range transport (e.g., Oerlemans et al., 2009; Dumont et al., 2014; Nagorski et al., 2019; Marshall and Miller, 2020). Actively deglaciating terrain typically exposes fresh sediment and rock surfaces at the glacier margins, some of which could be mobilised and deposited on the glacier. Oerlemans et al. (2009) suggest that this process may be causing an increase in dust concentrations and melt rates on the lower section of Morteratsch Glacier, Switzerland.

Organic material also constitutes a significant fraction of particulate matter on some glaciers, much of it associated with *in situ* algal populations (e.g., Takeuchi et al., 2001, 2006; di Mauro et al., 2020; Williamson, C. J. et al., 2020). There is an interaction between melting, mineral impurities, and biological activity, as algae and cyanobacteria require liquid water and nutrients. Warming temperatures, increase in liquid water, and longer melt seasons may be driving an increase in algal populations on glaciers and ice sheets. Albedo reductions due to algal activity have been reported at several sites (Stibal et al., 2017; Williamson et al., 2019; di Mauro et al., 2020; Williamson, C. J. et al., 2020).

Organic material and black carbon deposits on glaciers are also derived from incomplete combustion of fossil fuels, biomass burning, and forest fires (Ming et al., 2009; Keegan et al., 2014; de Magalhães Neto et al., 2019; Nagorski et al., 2019). Black carbon associated with industrial activity has been implicated in glacier darkening in the Himalayas (Ming et al., 2009, 2015). Albedo reductions have also been associated with the deposition of aerosols from wildfire activity (Keegan et al., 2014; de Magalhães Neto et al., 2019; Marshall and Miller, 2020). Wildfire represents a potential indirect effect of climate change on glaciers. There is a concern that increasing wildfire activity in western North America may be darkening glaciers in the region, contributing to increased glacier melt. Glaciers in the Canadian Rocky Mountains are vulnerable to this as they are situated downwind of wildfire activity in the U.S. Pacific Northwest and British Columbia. Marshall and Miller (2020) report ice-albedo values of as low as 0.07 during the summer of 2017, which was a record wildfire season in British Columbia at the time.

Regardless of its provenance, increased particulate matter is a positive feedback because it enhances melting, which concentrates impurities and further lowers the albedo. Indirect effects of climate change such as increasing wildfire and algal activity or increased dust deposition associated with glacier retreat exacerbate the melt-albedo feedback. While increasing concentrations of particulate matter are more of a gradual process than a regime change, the resulting melt-albedo feedbacks are accelerating the processes of deglaciation and the transition of the world's mountain regions from white to grey.

Seasonal Transition From Snow to Ice

Even in the absence of external darkening agents, climate change directly impacts surface albedo on glaciers. As noted above, the summer melt-season transition from seasonal snow to exposed glacier ice is attended by a sharp drop in surface albedo on most mountain glaciers. Reductions in winter snowpack, an earlier start to the melt season, and higher melt rates all lead to an earlier transition from seasonal snow to exposed glacier ice. In addition, transient increases in albedo from summer snow events has been shown to be important processes in reducing mass loss from the

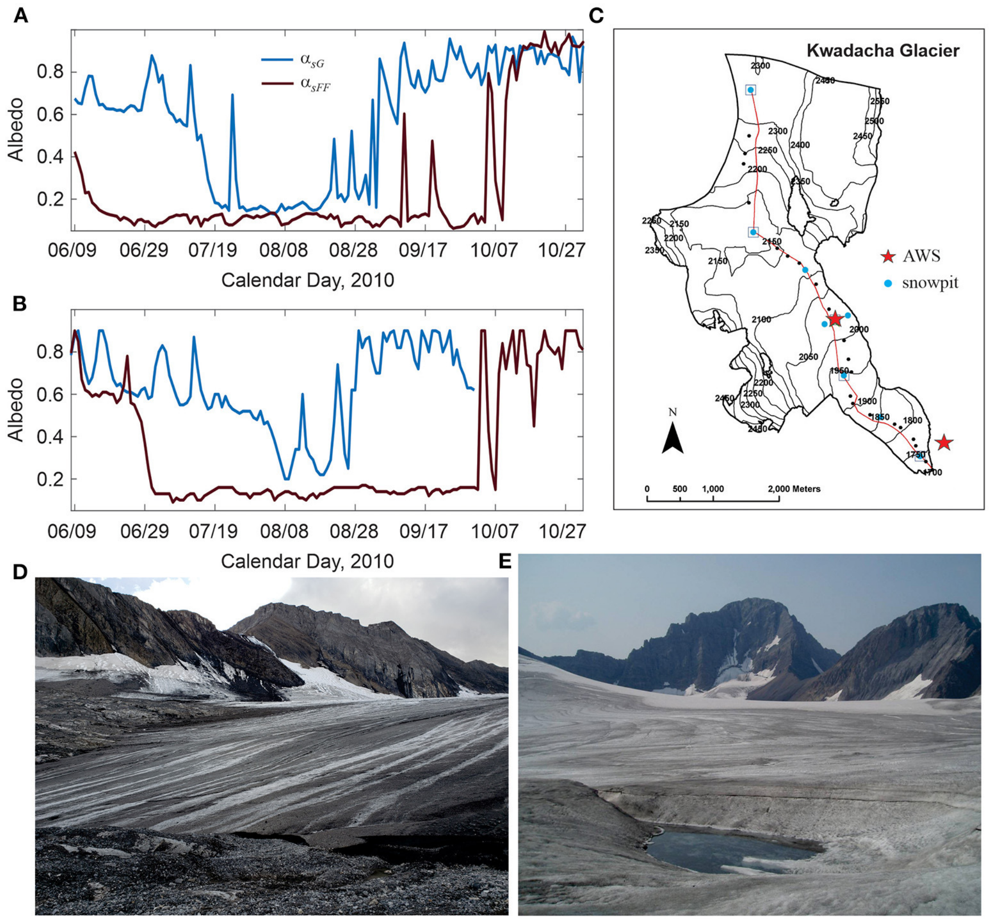

FIGURE 7 | Mean daily surface albedo through the summer melt season as measured by paired on- and off-glacier automatic weather stations at **(A)** Kwadacha and **(B)** Haig Glaciers in the Canadian Rockies. Both plots show the albedo evolution from June 8 to November 1, 2010 from glacier sites in the upper ablation area (blue) and from proglacial sites in the glacier forefields (brown). **(C)** Contour map of Kwadacha Glacier, with the stars indicating the AWS locations. Photographs **(D,E)** are August, 2021 and August, 2006 images of Haig Glacier, providing visual examples of exposed glacier ice with an albedo of ∼0.2 on the lower and upper glaciers. The end-of-summer seasonal snowline is visible in **(E)**. Photographs by the author.

glacier (Marshall and Miller, 2020). These summer "refreshes" of the glacier surface will be less frequent under warmer conditions, as precipitation on the glacier shifts from snow to rain.

Figure 7 plots examples of these influences for the summer 2010 melt season at two different mid-latitude mountains glaciers in the Canadian Rocky Mountains. Automatic weather station (AWS) pairs were established for multi-year periods at each site: one on the glacier, in the upper ablation area, and one in the glacier forefield region (cf. **Figure 7C**). The off-glacier AWS at each site is situated on bare limestone. Marshall (2014) describes the Haig Glacier AWS records in detail. At Kwadacha Glacier, the forefield AWS was set up at an elevation of 1,670 m in August, 2007 and recorded continuously until August, 2011. The glacier AWS was set up at the same location on the glacier each year from 2008 to 2011, at an elevation of 2,005 m, and was established between mid-May and early June, while the glacier surface was still snow-covered.

The two glaciers are about 1,010 km apart and are not subject to the same weather systems through the summer of 2010, but the melt-season albedo evolution is similar at the two locations (**Figure 7**). All four AWS sites were snow-covered in the first week of June. Snow at the off-glacier sites melted away by the end of June, exposing bare rock with an albedo of ∼0.1. These sites remained mostly snow-free until mid-October. At Kwadacha Glacier the transition from seasonal snow to bare glacier ice took place in the second week of July, with bare ice (albedo of ∼0.2) persisting until the first week of September (**Figure 7A**). Several summer snow events caused abrupt and transient increases in the glacier albedo, persisting for 1–3 days. Snow began to accumulate on the glacier in the second week of September, heralding the start of "winter" (i.e., the beginnings of the 2010–2011 winter snowpack). The Kwadacha forefield site experienced two transient September snow events in 2010, evidenced by short-lived albedo spikes (**Figure 7A**). These occurred after snow had already begun to accumulate on the glacier, and the main

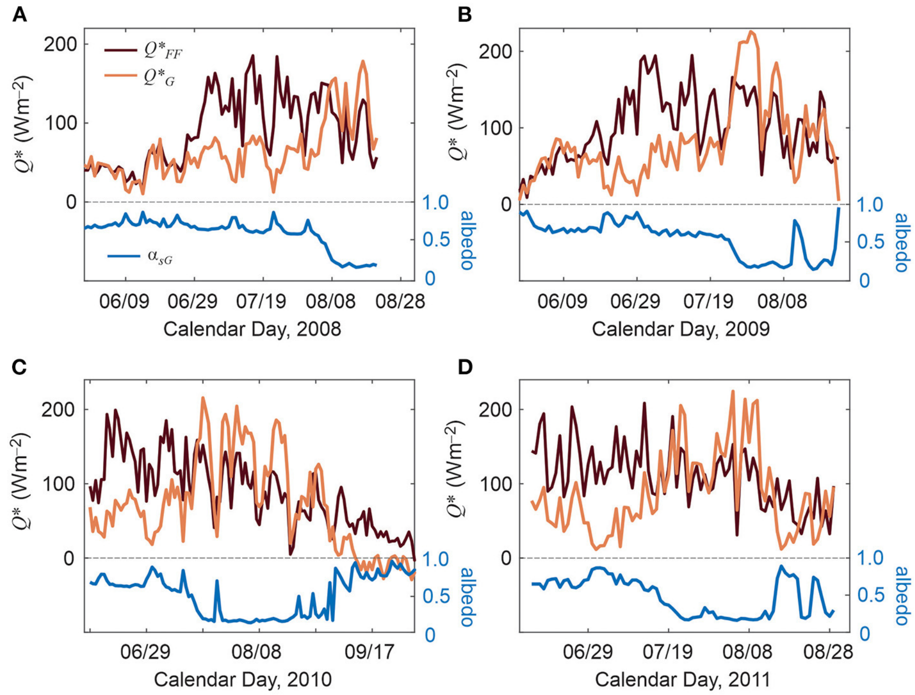

FIGURE 8 | Mean daily net radiation, Q*, at the glacier forefield AWS and the glacier AWS for the **(A)** 2008, **(B)** 2009, **(C)** 2010, and **(D)** 2011 melt seasons at Kwadacha Glacier. The rapid decrease in mean daily glacier surface albedo, plotted in blue indicates the transition from seasonal snow to exposed glacier ice each summer.

melt-season snow events on the glacier were not registered at the forefield AWS.

Summer albedo evolution at Haig Glacier was similar, with a drop in albedo from ∼0.55 to ∼0.2 marking the transition to exposed glacier ice in the first week of August (**Figure 7B**). Similar to Kwadacha Glacier, several summer snow events are evidenced by transient albedo spikes at the glacier AWS through the months of June to August. The bare-ice season was relatively short at Haig Glacier in 2010, with winter snow accumulation at the glacier AWS commencing on August 28. The Haig Glacier forefield AWS did not register any summer snow events in 2010, though there was one transient event in early October, before the off-glacier winter snowpack started to accumulate the following week (**Figure 7B**). **Figures 7D,E** provide a visual perspective of a glacier surface with an albedo of ∼0.2, for the lower ablation area and the upper accumulation area of Haig Glacier. On the lower glacier, the glacier ice is difficult to discern from the limestone rock and till. On the upper glacier, particulate matter and melt ponds can both contribute to the low-albedo surface. The late-summer snowline is visible at the highest elevations of the glacier, near the top of the image.

This seasonal albedo evolution is typical of mid-latitude mountain glaciers, and illustrates the strong changes that attend both (a) the transition from seasonal snow cover to exposed glacier ice, and (b) transient summer snow events, which temporarily refresh the glacier surface (Oerlemans and Klok, 2003). Each of these conditions strongly affect the glacier energy and mass balance. **Figure 8** illustrates this for four different summers on Kwadacha Glacier, 2008 to 2011, plotting the seasonal evolution of mean daily net radiation and albedo on the glacier. Net radiation is the sum of all incoming and outgoing shortwave and longwave radiation fluxes,

$$Q^* = Q_S{}^{\downarrow} + Q_S{}^{\uparrow} + Q_L{}^{\downarrow} + Q_L{}^{\uparrow} = Q_S{}^{\downarrow}(1 - \alpha_s) + Q_L{}^{\downarrow} - \varepsilon_s \sigma T_s{}^4, \tag{8}$$

for surface albedo α_s, emissivity ε_s, and Stefan-Boltzmann's constant σ. Net radiation is positive during the summer melt season, and increases ∼3-fold during the transition from seasonal snow to low-albedo glacier ice. This is the dominant term in the glacier surface energy balance, accounting for roughly 80% of the net energy available for glacier melt in these environments (e.g., Klok and Oerlemans, 2002; Marshall, 2014).

Net radiation from the Kwadacha forefield AWS site provides an interesting comparison with net radiation on the glacier (**Figure 8**). The AWS records from 2008 and 2009 start in mid-May (**Figures 8A,B**), when the glacier forefield site was still snow-covered. At these times, net radiation on and off the glacier is nearly identical. The elevation difference between these sites is 335 m, which has a minor and off-setting influence on incoming radiation fluxes. Differences in net radiation between the sites are therefore associated with the albedo and surface temperature, via Stefan-Boltzmann's relation for outgoing longwave radiation in Equation (1). Net radiation increases strongly at the forefield site once it is snow-free, and greatly exceeds the net radiation available on the glacier while the glacier remains snow-covered. On average over the four summers ($\pm$ 1 standard deviation), net radiation at the forefield site exceeds that on the glacier by 52 $\pm$ 39 Wm^{-2} when there is snow cover at the glacier AWS (**Table 2**).

The opposite is true when glacier ice is exposed; over bare ice, net radiation on the glacier exceeds that in the forefield by 36 $\pm$ 24 Wm^{-2} (**Table 2**). Counter-intuitively, this means that there is more net radiative energy available on the glacier than in the proglacial environment. This is partially driven by the increase in absorbed shortwave radiation over bare ice. The ice albedo is higher than that of the proglacial environment, however, so this is not the full storey. Surplus energy on the glacier is also due to the fact that the surface temperature cannot exceed the melting point, 0°C. This limits the outgoing longwave radiation to 315 Wm^{-2} on the glacier, whereas outgoing longwave radiation in the forefield increases as the rock warms up, averaging 353 Wm^{-2} over the summer.

Summer Snowfall

The net radiation "inversion" on the glacier prevails in the ablation zone, where surface albedo is not much greater than that of the proglacial rock or sediments. This situation temporarily reverses during summer snow events on the glacier (**Figure 8**, **Table 2**). For the August snow events on Kwadacha Glacier from 2008 to 2011 (17 days in total), the average daily net radiation is 33 Wm^{-2}, compared with 104 Wm^{-2} in the glacier forefield and 130 Wm^{-2} for bare-ice days in late summer. These summer snow events on the glacier, evident in the albedo records in **Figures 7**, **8**, all registered as rainfall in the glacier forefield.

Marshall and Miller (2020) assess the importance of summer snow events on Haig Glacier to total melt-season surface energy and mass balance. The impact varies from year to year, but on average, summer snow events increase the mean summer (JJA) albedo in the upper ablation area by about 0.07, from an estimated snow-free value of 0.48 to the observed mean value of 0.55. This results in a ~15% reduction in total summer melt and runoff. Summer snow events don't tend to involve much accumulation of mass, but their direct impact on albedo is significant to the glacier mass balance (Oerlemans and Klok, 2003). This effect is threatened by the likelihood that summer precipitation events will increasingly shift from snowfall to rainfall on glaciers. For the August snow events on Kwadacha Glacier from 2008 to 2011, the average daily forefield and glacier temperatures were 4.0 and 1.2°C, respectively. This is near the transition from liquid to solid-phase precipitation, where a slight warming will result in a shift from snow to rain on the glacier.

Multi-Decadal Perspective

The long-term evolution of surface albedo and net radiation can be modelled where AWS data or bias-adjusted climate reanalyses are available to drive a model of surface energy and mass balance. As an example, **Figure 9** presents a historical reconstruction of mean summer (JJA) surface albedo and net radiation along the central flowline of Kwadacha Glacier, from the upper accumulation area to the terminus. Results are calculated from a surface energy balance model driven by ERA5 climate reanalyses for the period 1950–2020. **Figure 9** illustrates the evolution of albedo and net radiation as a function of distance downglacier for the 1950s, 1980s, and 2010s.

The progressive decrease in surface albedo over these decades is evident, driven by the general trend of longer and warmer melt seasons. Averaged over the glacier, surface albedo has a trend of -0.012 per decade and net radiation has a trend of $+3.7$ W m^{-2} per decade from 1950 to 2020, i.e., changes of -0.08 and $+26$ W m^{-2} over this period. These changes are significant, relative to mean values ($\pm 1\sigma$) of 0.58 ± 0.05 and 72 ± 14 W m^{-2} from 1950 to 2020. The linear correlation coefficient between time series of mean glacier albedo and net radiation (1950–2020) equals $r = -0.98$, indicating that interannual variability in net radiation is primarily governed by the albedo. Increases in net radiation over this 71-year period are therefore dominated by the albedo feedback. The decadal-scale decrease in albedo and increase in melt energy is partially associated with the up-glacier propagation of the ablation area and the relatively recent exposure of bare glacier ice at higher elevations. The flowline data from the 2010s illustrate this effect, with the abrupt step at $x = 3.2$ km. This aligns with the average elevation of the end-of-summer snowline through this decade.

The model results in **Figure 9** do not include the potential effects of increased particulate loading or algal activity on the glacier, as these effects have yet to be interactively coupled in glacier mass balance models. The results are just due to the direct climate change effects of a longer and more spatially-extensive bare-ice season, along with a reduction in summer snow events. These direct albedo feedbacks are driving increased mass loss and accelerated glacier retreat (Ebrahimi and Marshall, 2016). Additional albedo reductions due to increased dust loading, wildfire or industrial activity, and algal development may be further hastening the transition from white to grey in some regions.

From White to Blue

Glacier and ice sheet retreat does not always expose bare rock. In polar regions at the ice-ocean interface, the collapse of floating glacier tongues and ice shelves represents a transition from white to blue. Where this is occurring, the loss of floating ice can have significant ecological impacts for local marine environments (e.g., Vincent and Mueller, 2020) and for oceanographic processes associated with ice-ocean interactions (e.g., mixing, stratification, sea ice and iceberg production, sedimentation). Catania et al. (2020) offer an excellent review of marine-terminating outlet glacier processes in Greenland and their ecological and oceanographic significance. Similar dynamics are impacting large sectors of the Antarctic Ice Sheet (e.g., Jenkins et al., 2018; Meredith et al., 2019).

Outlet glacier fluctuations at the ice sheet-ocean margin are complex and not fully understood (Catania et al., 2020). The inland retreat of tidewater glaciers and floating ice tongues causes short-term acceleration of ice sheet flow and higher rates of sea-level rise, as marginal thinning propagates upstream (Price et al., 2011; Catania et al., 2020; King et al., 2020). Following a transition period, however, outlet glacier margins may retreat to inland positions where they are more stable. In the absence of calving losses and marine melting at the ice-ocean interface,

TABLE 2 | Mean daily albedo and net radiation, Q^*, recorded at the glacier and off-glacier automatic weather stations on Kwadacha Glacier, 2008–2011.

		Forefield AWS		Glacier AWS		ΔQ*
Period	N_d	**albedo**	**Q* (W m^{-2})**	**albedo**	**Q* (W m^{-2})**	**(W m^{-2})**
Summer (JJA)	326	0.20 ± 0.21	104 ±43	0.51 ± 0.24	83 ±50	−21 ±54
Seasonal snow	208	0.26 ± 0.25	108 ±45	0.68 ± 0.10	56 ±26	−52 ±39
Glacier ice	113	0.10 ± 0.02	94 ±37	0.20 ± 0.04	130 ±50	36 ±24
August snow	17	0.08 ± 0.01	104 ±43	0.70 ± 0.13	33 ±18	−44 ±25

Mean values (± 1 standard deviation) are shown for all days with available summer (JJA) observations from both stations. N_d is the number of days with observations within each period and ΔQ^* is the difference in net radiation on the glacier, relative to the glacier forefield. "Full summer" shows the values for all days, "seasonal snow" shows the subset of days with seasonal (winter) snow at the glacier AWS, primarily in June and July, "glacier ice" is the subset of days with exposed glacier ice at the glacier AWS, primarily in August, and "August snow" is for all days when summer snow events transiently covered exposed glacier ice, as recorded at the glacier AWS.

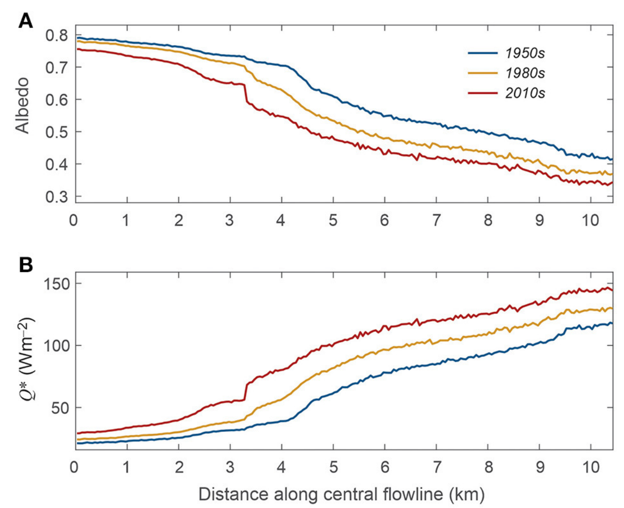

FIGURE 9 | Mean summer (JJA) **(A)** albedo and **(B)** net radiation, Q^*, along the Kwadacha Glacier central flowline for 1950–1960 (blue), 1980–1990 (orange), and 2010–2020, based on surface energy balance modelling from 1950 to 2020. The top of the flowline ($x = 0$) extends to Mt. Lloyd George, at an elevation of 2,938 m, and the terminus elevation ($x = 10.4$ km) is ∼1,700 m.

outlet glaciers that have lost touch with the ocean will have a more positive mass balance. This is one of the few negative feedbacks associated with glacier retreat processes, although it will require decades to centuries for some of the major outlets of the Greenland Ice Sheet to retreat to a stable terrestrial terminus position. Such a transition may help to reduce the rate and magnitude of sea-level rise from the Greenland Ice Sheet in the coming centuries, but marine-terminating outlet glaciers that have retreated inland will still be susceptible to ongoing atmospheric warming (Aschwanden et al., 2019).

This process of "self-stabilizing" glacier retreat is in fact common for mountain glaciers, which can retract to higher altitudes with less negative mass balance. This is well-manifest in north-facing cirque glaciers that have retreated to a topo-climatic niche where they are relatively stable. This can be thought of as a regime change from negative to neutral or positive mass balance. Glaciers can weather short-term warming through this process, but as in the polar regions, ongoing warming will make it difficult to reach a new and sustainable equilibrium.

In terrestrial environments, glacier retreat often exposes bedrock basins that have been over-deepened by glacial erosion, creating proglacial lakes in the wake of the retreating ice (e.g., Shugar et al., 2020). Proglacial lakes often pose a hazard, due to the risk of outburst floods (Carrivick and Tweed, 2016; Huggel et al., 2020), but they can also serve to moderate the hydrological impacts of diminished glacier ice, by acting as reservoirs to store water and release it slowly through the summer. Ice-marginal proglacial lakes also increase glacier mass loss through calving (Benn et al., 2007; Sakai et al., 2009; Maurer et al., 2016). This is another positive feedback process that can contribute to accelerated glacier mass loss (e.g., King et al., 2019).

From Sublimation to Melting

Meteorological conditions that are favourable to sublimation of snow and ice (high winds, low humidity, sub-zero temperatures) focus available sensible and radiative energy on sublimation rather than melting (e.g., Wagnon et al., 1999; Kaser, 2001; Mölg et al., 2008; MacDonell et al., 2013). Sublimation can serve as an effective energy sink, consuming significant amounts of net energy with minimal mass loss; compared with melting, it requires 8.5 times more energy to sublimate a molecule of ice (Cuffey and Paterson, 2010). This makes sublimation an ineffective ablation mechanism on glaciers, resulting in a more positive mass balance. This is well-documented for high-altitude glaciers in the Himalaya, the tropical Andes, and East Africa (e.g., Mölg et al., 2008; Ayala et al., 2017), and similar processes are important on the Greenland and Antarctic Ice Sheets.

Tropical glaciers spanning a large elevation range commonly experience a gradation in the dominant ablation regime from melting to sublimation as one moves to altitudes above ∼5,000 m (e.g., Ayala et al., 2017). Cold and dry conditions that favour sublimation at high altitudes are sensitive to increases in both temperature and atmospheric moisture that are occurring with climate change (Trenberth et al., 2005; Hartmann et al., 2013; Gao et al., 2018; Ho et al., 2018). These changes increase incoming longwave radiation and latent heat flux in glacial environments, and could precipitate a transition from sublimation to melting at high elevations, non-linearly increasing ablation and mass loss. The partitioning of ablation processes between sublimation

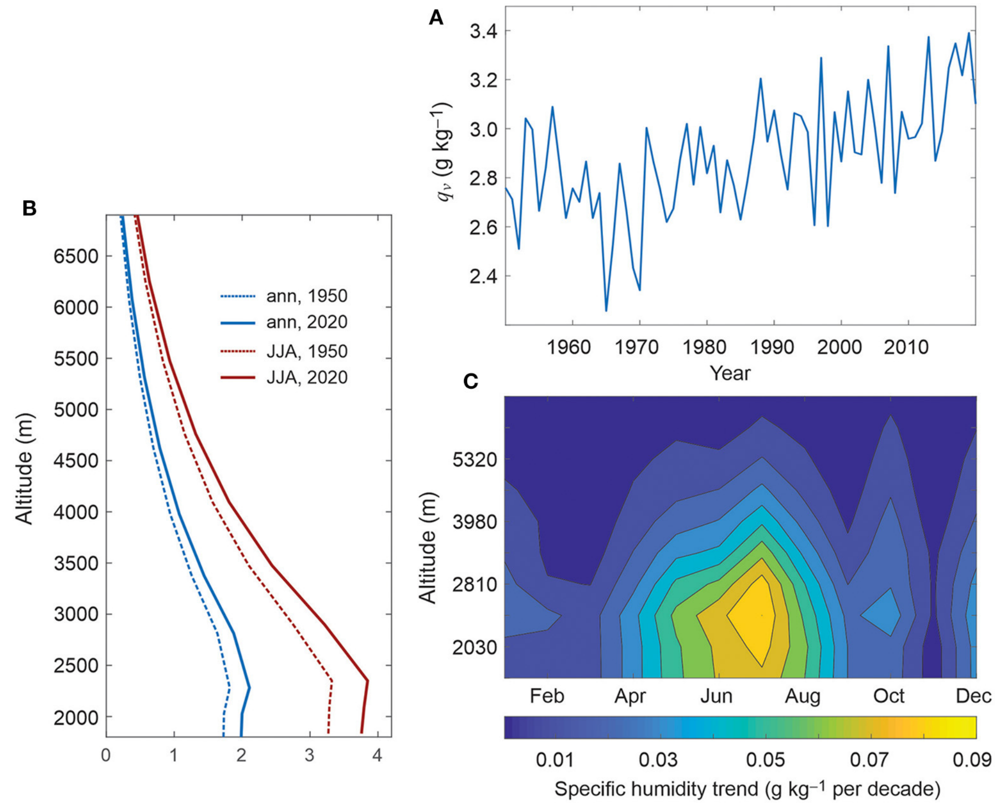

FIGURE 10 | Specific humidity evolution from 1950 to 2020 in the ERA5 climate reanalysis in the grid cell over the upper Kaskawulsh Glacier, St. Elias Mountains, Yukon. **(A)** Mean annual specific humidity at the 700-mb level (∼2,810 m). **(B)** Profiles of mean annual and summer (JJA) specific humidity in the 1950s (dashed lines) and the 2010s (solid lines). **(C)** Specific humidity trends over the period 1950–2020 as a function of pressure-level (altitude) and month.

and melt is critical to glacier response to climate variability and change in high-elevation tropical environments (e.g., MacDonell et al., 2013), and shifts from sublimation to melt have been identified as a potential driving mechanism for glacier retreat on Mt. Kilimanjaro (Thompson et al., 2009) and in the tropical Andes (Réveillet et al., 2020).

Williamson S. et al. (2020) report positive trends in atmospheric humidity as the main manifestation of elevation-dependent climate change in the St. Elias Mountains of Yukon Territory, Canada. Climate reanalyses indicate marked increases in humidity at high elevations in this region, related to moisture advection from the North Pacific. **Figure 10** shows an example of this from the pressure-level specific humidity data in the ERA5 grid cell over the Kaskawulsh Glacier field site (cf. **Figure 2**). Specific humidity at 700 mb (∼2,810 m altitude) increased by about 14% from 1950 to 2020 (**Figures 10A,B**), with the strongest absolute changes in the summer months (JJA) and at the 700-mb level (**Figure 10C**). As a percentage, annual and summer changes are similar, with an average increase of 1.9% per decade over the pressure levels from 400 to 800 mb plotted in **Figures 10B,C**. Increases in atmospheric specific humidity reduce humidity gradients at the glacier surface, lowering sublimation rates and the energy sink association with latent heat flux. Together with influences on incoming longwave radiation, this is an important consideration for analyses of elevation-dependent climate change, with potential for non-linear impacts on glacier mass balance.

DISCUSSION

Processes Governing Non-Linear Glacier and Ice Sheet Response to Climate Change

The transitions discussed above are all under way as part of glacier and ice sheet response to climate change, but they are non-contemporaneous in space. There will not be a globally synchronous switch from polythermal to temperate ice masses or from meltwater retention to runoff, net accumulation to ablation, snow to rain, sublimation to melting, or white to grey. These transitions can nonetheless be rapid at any one point, representing a veritable regime shift.

Because these are non-linear, threshold processes, they can be difficult to predict with confidence in numerical models. As an example, the modelled transition from polythermal to temperate firn at Kaskawulsh Glacier in **Figure 2** was an unanticipated model result, which may or may not reflect reality. There is temperate, water-saturated firn at 35 m depth at the site at present, discovered during drilling of a firn core in 2018, but the age of this firn aquifer is unknown. The modelling predicts that this developed over the past decade in response to increasing surface melt at the site, with high amounts of meltwater infiltration in 2013 exceeding the refreezing capacity of the firn. This triggered the rapid transition to temperate conditions. The atmospheric warming signal at this site is gradual and subtle, however. Mean annual air temperatures averaged -11.2°C from 1950 to 2020, with a long-term trend of $+0.28^{\circ}$C per decade and a mean value of -10.3°C from 2010 to 2020. Mean summer (JJA) temperatures averaged -2.8°C, with a long-term trend of $+0.21^{\circ}$C per decade and a mean value of -1.8°C from 2010 to 2020. Conditions in the last decade were therefore only $\sim 1^{\circ}$C warmer than the long-term average. The coupled surface energy balance, firn thermodynamic, and firn hydrological system translated a gradual, linear atmospheric temperature forcing into the strongly non-linear response shown in **Figure 2**.

Latent heat release at this site is sufficient to produce temperate firn and ice despite mean annual air temperatures of about -10°C, but it is difficult to determine a threshold temperature which can trigger the transition from polythermal to temperate conditions. The regime change at Kaskawulsh Glacier was sudden, but the preconditioning for this transition developed over a number of years. Similar processes are seen in the modelled development of deep temperate firn at DYE-2, Greenland in **Figure 5**. Firn warming is gradual through the period 2000–2077, though at about twice the rate of the atmospheric warming, until a warm summer with above-average surface melting in 2077 triggers the transition to deep temperate conditions (**Figures 5B,C**). The years immediately preceding this pre-conditioned DYE-2 for this transition (**Figure 5E**). Following the transition to deep temperate conditions, air temperatures and melt rates at DYE-2 are high enough to sustain temperate firn.

This transition to temperate conditions at DYE-2 does not occur with the reference climatology of **Figure 4**, however; it requires significantly ($> 40\%$) higher snow accumulation rates at DYE-2. It is more likely that this site will transition to an ablation area without developing temperate conditions in response to projected climate warming.

Latent heat of refreezing is the main factor that drives the development of temperate conditions and the magnitude of the difference between 10-m firn temperature and mean annual air temperature. In the simulations presented here, this is governed to first order by the amount of melt, and increasing melt rates provide the trigger for temperate firn development. It can develop abruptly and is difficult to reverse because once meltwater infiltrates to depths of $> \sim 8$–10 m, it is below the depth of the annual winter cold wave. Wet and temperate conditions develop if deep firn has insufficient cold content to refreeze all of the meltwater. It would require sustained climate cooling over many years to restore sub-freezing conditions, limited by time scales of thermal diffusion from the surface. The latent heat release that drives this transition can be suppressed if the seasonal snowpack and near-surface firn do not have sufficient cold content, pore space, and permeability to support meltwater infiltration and refreezing. In particular, the development of thick, near-surface ice layers can inhibit meltwater infiltration (Bezeau et al., 2013; MacFerrin et al., 2019), potentially diverting meltwater to lateral runoff. There are numerous uncertainties in the firn system, particularly meltwater infiltration processes and the formation and permeability of ice layers (van As et al., 2016; Samimi et al., 2020; Vandecrux et al., 2020b).

Non-linear processes in glacier surface energy and mass balance are similarly hard to predict. Temperature strongly modulates multiple feedback processes that contribute to glacier response to climate change, including shifts from snowfall to rainfall, declines in glacier albedo, the magnitude and sign of latent heat fluxes, and the extent of meltwater refreezing. These interact to increase glacier sensitivity to warming beyond the direct effects of increases in sensible and incoming longwave radiation fluxes, which scale with temperature. These interactions help to explain the strong global glacier response to a warming of about 1°C over the past four decades (Marzeion et al., 2014; Hugonnet et al., 2021). Ebrahimi and Marshall (2016) analysed the sensitivity of glacier mass balance to a warming of 1°C at Haig Glacier with and without mass balance feedbacks, using an energy balance model calibrated to observations of glacier melt. For the period 2002–2012, the average summer (JJA) net energy and summer melt at the glacier AWS site were $Q_N = 97$ W m^{-2} and $\dot{m} = 2.32$ m w.e. yr^{-1}. A temperature change of $+1^{\circ}$C, with feedbacks suppressed, gives $\Delta Q_N = 8.3$ W m^{-2} and a 9% increase in summer melt ($\dot{m} = 2.52$ m w.e. yr^{-1}). Including humidity and albedo feedbacks in the surface energy balance, a 1°C-warming gives $\Delta Q_N = 27$ W m^{-2} and $\dot{m} = 2.97$ m w.e. yr^{-1}, a 28% increase in summer melt.

This is consistent with other studies on mid-latitude mountain glaciers, which report a $\sim 30\%$ reduction in net mass balance for a warming of 1°C (Klok and Oerlemans, 2002; Hirose and Marshall, 2013; Huss and Fischer, 2016). Feedbacks increase glacier mass balance sensitivity to a change in temperature by a factor of ~ 3. This implicitly includes some albedo feedbacks, such as a longer melt season with a greater duration of exposure of low-albedo glacier ice and more frequent summer rain events rather than snowfall as temperatures warm. This estimate does not include the potential effects of inheritance of impurities over many years, which can increase in concentration as meltwater runoff leaves some of the particulate load behind on the glacier surface. Additional albedo reductions that are not captured include the possible influences of increased glacier algal activity, dust loading from recently-deglaciated terrain, or increases in black carbon deposition. At Haig Glacier, for instance, the average long-term bare-ice albedo at the AWS site is 0.21 ± 0.06, but values of ~ 0.1 were measured for several weeks in August 2017, associated with intense wildfire activity upwind of the glacier (Marshall and Miller, 2020). Mean measured August albedo at the AWS site was 0.15 in 2017, compared with a long-term average of 0.38 (Ebrahimi and Marshall, 2016). This albedo reduction over the month of August equates to an additional melt of 0.38 m w.e., or a 16% increase in total summer runoff. Any additional darkening from increased particulate matter therefore serves to further increase the climate change sensitivity of ice masses.

Other direct consequences of the transition from white to grey or blue (exposed rock, melt ponds, and proglacial lakes) include sensible and radiative heat fluxes from exposed rock and open water. Glacier calving into marginal lakes can further accelerate glacier retreat (Watson et al., 2020). Observed increases in glacier mass loss over the past two decades testify to the current state of disequilibrium (Hugonnet et al., 2021).

Considerations for Glacier-Climate Modelling

Many of the feedback processes of glacier melt and retreat are not well-captured in glacier-climate models, which typically have one-way climate forcing and do not generally simulate the evolving environment (e.g., changes in proglacial terrain, lake development, ice-ocean conditions, transport, and deposition of impurities). Because many of these processes involve positive feedbacks, current model projections may underestimate glacier and ice sheet response to climate change. Process studies to advance understanding and modelling of some of these non-linear glacier mass balance processes can increase confidence in future projections.

It is important to understand mechanisms of elevation-dependent climate change and how these interact with glacier-climate processes on both mountain glaciers and polar ice sheets. Changes in the phase of precipitation with altitude, lapse rates of temperature, specific humidity, and longwave radiation, and overnight meltwater refreezing processes are all important to glacier surface energy balance, and will evolve in poorly-understood ways in response to ongoing climate warming. Coupled evolution of the glacier-climate system and explicit representation of the surface energy balance (vs. temperature index or degree-day melt models) may better capture many of these processes and feedbacks. An Earth-systems approach is also needed to represent the broader evolution of glacier, lake, ocean, and proglacial alpine/tundra environments as glaciers retreat from the landscape. This is important to regional energy fluxes and climate, as well as ecological and hydrological conditions and emerging hazards in deglaciating terrain.

While these processes are non-linear and can result in abrupt regime changes, it is difficult to identify the glaciological or glacio-climatic tipping points with confidence. The evolution from polythermal to temperate conditions and warming of the ice in glaciers and ice sheets may play a significant role in long-term ice-dynamical response to climate change, but the evolution of firn in Greenland and in parts of Antarctica which likely to experience surface melt this century is uncertain. Firn simulations like that shown in **Figures 2–5** can be thought of as physically-plausible scenarios rather than predictions. These results are sensitive to uncertain process representations and parameter settings in the firn model, in particular concerning mechanisms of meltwater infiltration, capillary meltwater retention (irreducible water content in the pore space), and ice-layer formation (Samimi et al., 2020). These processes need to be better understood and represented in models before non-linear thresholds within firn temperature evolution and meltwater retention can be reliably predicted.

Ensemble modelling (multiple realisations) spanning a range of parameter space (e.g., Van Pelt et al., 2021) can arguably offer reasonable projections of how firn may evolve in response to climate change. For instance, firn conditions and meltwater runoff to retention ratios for different parts of the Greenland Ice Sheet could be expressed via probability distributions of conditions for the period 2070–2100 for a given climate scenario. This is a reasonable approach to future projections at this time, although averaging within ensemble modelling artificially smooths the projections and make regime changes appear like gradual system responses, rather than the rapid transitions that are expected in nature. Structural uncertainties within firn hydrological process models (Vandecrux et al., 2020b) may also bias ensemble modelling, giving distributions that don't fully represent the range of future conditions.

CONCLUDING THOUGHTS

This review is not comprehensive in its identification of threshold processes and regime changes which glaciers experience as they respond to gradual climate forcing. In particular, several elements of glacier and ice sheet dynamics are non-linear and can either accelerate or help to stabilise glacier retreat. Marine grounding-line and tidewater glacier instabilities are two well-known examples of this, where an unstable state of retreat can be triggered for thinning ice on a reverse bed slope (e.g., Weertman, 1974; Schoof, 2007; Joughin et al., 2014). Climate-driven thinning at the margin of grounded ice masses also steepens glacier slopes, increasing flow rates and advection of ice to the ablation zone (e.g., Huybrechts and De Wolde, 1999; Schäfer et al., 2015), a positive feedback that accelerates mass loss. In other situations, retreating ice masses can stabilise inland or at higher elevations, if the climate permits a glacier to establish a new equilibrium with a neutral rather than negative mass balance.

The transitions from polythermal to temperate firn and ice, from meltwater retention to runoff, from solid to liquid phases of water (i.e., snow to rain; reduced overnight refreezing of meltwater), from bright snow to low-albedo ice surfaces and deglaciated terrain, and from sublimation to melting-dominated ablation at high altitudes all represent regime shifts within the glacier-climate system. These involve non-linear processes that accelerate glacier and ice sheet response to climate change once external forcing passes a particular threshold. Gradual forcing can translate to an abrupt change in system response or behaviour. These regime changes are not intrinsically irreversible, but hysteresis and inertia in the system mean that the original states can be difficult to restore. Latent heat release from meltwater refreezing immediately warms firn and ice (e.g., **Figure 2**), but thermal diffusion is the only mechanism for cooling and restoring polythermal conditions, on decadal timescales for deep firn. Similarly, glacier and ice sheet surfaces with high concentrations of particulate matter will remain dark until these advect through the system to the glacier margin. Alternatively, climate cooling could allow these surfaces to more effectively covered by reflective snow for a greater proportion of the melt season.

Stabilisation of climate change is not likely to restore polythermal conditions, refresh glacier surfaces, or halt glacier retreat, due to the current state of disequilibrium. Hence, while glaciers and ice sheets are not past a "tipping point" that guarantees their complete disappearance, it will be difficult to stop or reverse their demise in the face of ongoing climate change (e.g., King et al., 2020; Hugonnet et al., 2021). The non-linear feedbacks identified in this discussion are not fully understood or adequately modelled, but are likely to contribute to accelerated glacier and ice sheet response to climate change in the coming decades. Process studies and model development to better represent these processes in Earth system models will enable improved projections of future change and a better representation of the specific climate thresholds that will accelerate such changes.

SUMMARY

Non-linear processes and feedbacks are contributing to the strong sensitivity of glacier and ice sheets to climate warming. This includes a number of threshold processes that can be triggered by a gradual climate forcing, such as regime shifts from polythermal to temperate glacier states and from snowfall to rainfall, sublimation to melting, and meltwater refreezing to runoff. Additional feedback mechanisms that can accelerate glacier mass loss include surface albedo reductions, increased calving losses, and increased heat flux from surrounding deglaciated terrain. These processes are not strictly irreversible but they create inertia in glacier retreat, such that it will be difficult to reverse the current state of disequilibrium in mountain glaciers and polar icefields. Many of the threshold processes involved in glacier-climate regime shifts are not well-understood and are still under development in modelling efforts. This includes meltwater routing and refreezing in firn and glacier systems, physically-based models of glacier surface albedo, and the evolving ice-atmosphere-land surface interactions in mountain and polar landscapes. This article identifies several regime shifts that are contributing to the extreme sensitivity of glaciers to climate change, which need to be better understood and modelled to permit improved projections of glacier change in the coming decades.

DATA AVAILABILITY STATEMENT

The datasets presented in this study can be found in online repositories. The names of the repository/repositories and accession number(s) can be found at: Datasets and model code used in this study are publicly available in the Scholars Portal Dataverse data repository at the University of Calgary. Haig and Kwadacha Glacier AWS data are archived at https://doi.org/10.5683/SP2/7CXPPI and https://doi.org/10.5683/SP2/QIPI9W, respectively. DYE-2 AWS and firn hydrology data are available at https://doi.org/10.5683/SP2/2QY39K. The MATLAB code for the surface energy balance and firn modelling is available at https://doi.org/10.5683/SP2/WRWJAZ.

AUTHOR CONTRIBUTIONS

The author confirms being the sole contributor of this work and has approved it for publication.

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Persistent Uncertainties in Ocean Net Primary Production Climate Change Projections at Regional Scales Raise Challenges for Assessing Impacts on Ecosystem Services

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Ocean net primary production (NPP) results from CO_2 fixation by marine phytoplankton, catalysing the transfer of organic matter and energy to marine ecosystems, supporting most marine food webs, and fisheries production as well as stimulating ocean carbon sequestration. Thus, alterations to ocean NPP in response to climate change, as quantified by Earth system model experiments conducted as part of the 5th and 6th Coupled Model Intercomparison Project (CMIP5 and CMIP6) efforts, are expected to alter key ecosystem services. Despite reductions in inter-model variability since CMIP5, the ocean components of CMIP6 models disagree roughly 2-fold in the magnitude and spatial distribution of NPP in the contemporary era, due to incomplete understanding and insufficient observational constraints. Projections of NPP change in absolute terms show large uncertainty in CMIP6, most notably in the North Atlantic and the Indo-Pacific regions, with the latter explaining over two-thirds of the total inter-model uncertainty. While the Indo-Pacific has previously been identified as a hotspot for climate impacts on biodiversity and fisheries, the increased inter-model variability of NPP projections further exacerbates the uncertainties of climate risks on ocean-dependent human communities. Drivers of uncertainty in NPP changes at regional scales integrate different physical and biogeochemical factors that require more targeted mechanistic assessment in future studies. Globally, inter-model uncertainty in the projected changes in NPP has increased since CMIP5, which amplifies the challenges associated with the management of associated ecosystem services. Notably, this increased regional uncertainty in the projected NPP change in CMIP6 has occurred despite reduced uncertainty in the regional rates of NPP for historical period. Improved constraints on the magnitude of ocean NPP and the mechanistic drivers of its spatial variability would improve confidence in future changes. It is unlikely that the CMIP6 model ensemble samples the complete uncertainty in NPP, with the inclusion of additional mechanistic realism likely to widen projections further in the future, especially at regional scales. This has important consequences for assessing ecosystem impacts. Ultimately, we need an integrated mechanistic framework that considers how NPP and marine ecosystems respond to impacts of not only climate change, but also the additional non-climate drivers.

Keywords: climate change, ocean net primary production, earth system model (ESM), climate projections, ocean modeling, oceanography, ocean biogeochemical cycles, ocean biogeochemical model

IMPORTANCE OF OCEAN NET PRIMARY PRODUCTION TO OCEAN ECOSYSTEM SERVICES

Photosynthesis by marine phytoplankton drives ocean net primary production (NPP, defined usually as photosynthesis minus respiration), which is a major planetary flux, accounting for half of global-scale photosynthesis (Field, 1998) and supporting major ocean ecosystem services. As a *de novo* source of fixed organic carbon, NPP primes the biological carbon pump and carbon sequestration, as well as supporting ecosystem structure and function (Falkowski et al., 2003; Bindoff et al., 2019). As such, changes to NPP in response to changes in climate act alongside other direct climate impacts, such as warming, to catalyse the response of these key services (Cheung et al., 2009; Lotze et al., 2019; Boyce et al., 2020; Tagliabue et al., 2020).

The net consumption of inorganic carbon and the production of particulate organic carbon during NPP perturbs the air-sea CO_2 partial pressure gradient, influencing atmosphere ocean carbon dioxide exchange. Subsequent transfer of this particulate organic carbon into the ocean interior *via* an array of particle injection pumps sequester carbon away from the atmosphere over various timescales (Boyd et al., 2019) and thus has an acknowledged major role in the global carbon cycle (Volk and Hoffert, 1985; Ito and Follows, 2005; Kwon et al., 2009). On the timescales of multiple centuries NPP acts to modulate atmospheric carbon dioxide levels *via* its control on the biological carbon pump intensity and ocean carbon inventories (Falkowski et al., 2003; Ito and Follows, 2005).

NPP is essential for marine biodiversity and associated ecosystem services. It is one of the main inputs of nutrients and energy that supports consumer growth and survival in marine food webs. Consumer diversity and their ecosystem functions then deliver important ecosystem services and societal benefits (Bindoff et al., 2019). These benefits include fisheries, climate regulation, supporting cultural, and intrinsic values. As an example, NPP is closely related to the distribution and magnitude of marine fisheries catches (Stock et al., 2017), which provide food, nutrition, income, and livelihoods for many millions of people globally (FAO, 2018). Changes in the transfer of NPP-derived particulate organic carbon from the ocean surface to the seafloor also affects the food supply to benthic ecosystems (Yool et al., 2017).

To achieve high rates of NPP, phytoplankton require light, and a range of resources for enzyme cofactors, as well as the structural components that permit high population levels (Falkowski et al., 1998; Saito et al., 2008). Alongside variable light conditions and the response of metabolic rates to temperature, deficiencies in the availability of these essential resources are the main "bottom up factors" limiting rates of NPP and driving its response to climate change (Laufkötter et al., 2015). Key resources to consider are nitrogen and iron, with phosphorus, silicon, and others playing roles in particular regions or for particular phytoplankton groups (Moore et al., 2013). Importantly, as the overall rates of NPP are affected by the combination of specific rates and population levels, they are also controlled by changes in predation by zooplankton (known as grazing) that affect phytoplankton standing stocks.

The Intergovernmental Panel on Climate Change (IPCC) made a range of assessments regarding NPP and associated ecosystem services in the recent Special Report on Oceans and Cryosphere in a Changing Climate (SROCC) report. In the SROCC, and other IPCC assessment reports, various different degrees of specific language are carefully used to assign confidence and likelihood to the assessments (Mastrandrea et al., 2011). Confidence is assigned based on the strength of the evidence and the level of agreement, with *high confidence* reserved for high agreement across robust evidence. Likelihood is used when enough confidence exists for there to be a probabilistic assessment, with *likely* reflecting 66–100% probability, *very likely* reflecting 90–100% probability and *virtually certain* related to 99–100% probability. When the results across multiple models are assessed, it is common to examine the *likely* or *very likely* range, either by using one or two standard deviations, respectively, or the appropriate confidence intervals (Collins et al., 2013; Bindoff et al., 2019).

NPP projections are available as part of the 5th and 6th Coupled Model Intercomparison Projects (CMIP5 and CMIP6, respectively). Due to the timing, SROCC relied on CMIP5 datasets and made the following assessments regarding the contemporary trends in NPP and future projections under the high emissions RCP8.5 scenario (Bindoff et al., 2019): (i) there was *high confidence* that the ongoing changes to stratification and nutrient cycles was having a regionally variable impact on NPP, (ii) there was *low confidence* assigned to the *very likely* range of NPP decline between 4 and 11% by 2081–2100, relative to 2006–2015, with *medium confidence* associated with projected declines of 7–16% (*very likely* range) in the tropical ocean that were constrained by historical variability. The projected declines in NPP were also assessed to lead to alterations of community structure (*high confidence*), reduce global marine animal biomass (*medium confidence*), and the maximum catch potential of fisheries (*medium confidence*), which may elevate the risk of impacts on income, livelihood, and food security of the dependent human communities (*medium confidence*). Taken together, these raise significant challenges for ocean ecosystem services in the face of ocean NPP changes.

PROCESSES GOVERNING OCEAN NET PRIMARY PRODUCTION AND THEIR REPRESENTATION IN EARTH SYSTEM MODELS

Generally, the types of ocean model used for climate change projections as part of CMIP6 and CMIP5 exercises follow a relatively similar approach to representing NPP despite showing a range of model configurations. They consider a discrete set of phytoplankton and zooplankton functional types (in rare cases also heterotrophic bacteria) and represent NPP as the product of the phytoplankton growth rate and biomass, with some inter-model differences in the representation of carbon loss due to respiration. In rare cases, gross primary production (GPP) is computed, with NPP the result of GPP minus autotrophic respiration and other loss terms (Vichi et al., 2007).

Phytoplankton growth rates are controlled by limitation by light and a range of nutrients using a Liebig "law of the minimum" approach. Across different models, these range from single nutrient limitation by N or P to accounting for multiple limiting factors (N, P, Si, and Fe). Fractional limitation of temperature dependent maximum growth rates is based on either external or cellular nutrient levels and when more than one nutrient is considered, the "most" limiting nutrient is used. Phytoplankton biomass is grazed by zooplankton, as a function of prey density following classical functional responses. The nutrient levels themselves are an emergent property of the ocean physical model and the balance between consumption during NPP and other processes, as well as the release and recycling by zooplankton, bacteria and degradation of sinking particulate organic matter, and sediment remineralisation in some models, which operate alongside external sources to the ocean (see Laufkötter et al., 2015; Séférian et al., 2020 for a more detailed description of NPP parameterizations in CMIP5 and CMIP6 Earth system models).

Earth system models show good skill in reproducing the distribution of major nutrients (nitrate, phosphate and silicate) (Séférian et al., 2020), with the notable exception of iron (Tagliabue et al., 2016), but less skill for indices of biological activity, like surface chlorophyll (Séférian et al., 2020). This issue is compounded for NPP, since the different types of observational algorithms and their various dependencies introduce additional uncertainty (beyond that for chlorophyll) in remote sensing estimates e.g., (Sathyendranath et al., 2020). Moreover, intercomparisons compare full Earth system models and thus the impact of differences in the specific NPP closures cannot be isolated from parallel differences in the accompanying ocean physical models and, more so, their atmospheric and land components. To date, the regional skill of Earth system models against estimates of NPP derived from remote sensing algorithms has not been conducted in depth, with global assessments available (Séférian et al., 2020).

Model projections of NPP at regional scales under different emissions scenarios are important because they inform community-level assessments, such as those by the IPCC, and form the basis of risk assessments for critical open ocean marine ecosystems and ecosystem services, such as fisheries (Bindoff et al., 2019). To date, uncertainty in NPP projections conducted as part of CMIP5 and CMIP6 at regional scale has been typically assessed by averaging changes across multiple years around the end of the century (2081–2100), relative to the recent historical era (1996–2015 for CMIP6) across models and considering whether more than 80% of the models agreed on the sign of change or with the statistical significance at regional scale (Bopp et al., 2013; Kwiatkowski et al., 2020). As a more quantitative assessment that is in line with IPCC assessments, the standard deviation around the multi-model mean can be used as the "*likely*" uncertainty range (Bindoff et al., 2013; Collins et al., 2013).

From the initial analysis of 13 CMIP6 Earth system models published by Kwiatkowski et al. (2020), global rates of NPP are projected to decline under the SSP5-8.5 high emissions scenario by a multi-model mean of 2.99% by the end of the 21st century, with a "likely" uncertainty range of ±9.1%. This spread has increased from the 10 models assessed for CMIP5 under the RCP8.5 scenario (Bopp et al., 2013), where there was a multi-model mean decline of 8.54%, with a "likely" uncertainty range around one-third lower at ±5.9%. The only regions where more than 80% of the CMIP6 models agreed on the sign of change under SSP5-8.5 were for the projected declines in the North Atlantic and increases in the Southern Ocean, with agreement on this metric more common for CMIP5 (Bopp et al., 2013). There is to date, no quantitative assessment of the regional uncertainty in absolute terms from the more recent CMIP6 models. Some efforts to assess regional change were performed as part of CMIP5 (Cabré et al., 2014; Laufkötter et al., 2015; Leung et al., 2015; Fu et al., 2016; Rickard et al., 2016; Kwiatkowski et al., 2017; Nakamura and Oka, 2019), highlighting the role of uncertainty and distinct regional drivers. Inter-model spread in the magnitude of NPP projections was noted in CMIP5 and was found to be localised in the tropics, but the regional expression and consequences for ecosystem services were not examined in great detail (Laufkötter et al., 2015).

As marine ecosystem projections are driven by changes in NPP or planktonic biomass, any inter-model differences and uncertainty in NPP projections then feeds into projected risk assessments under different climate change scenarios. For instance, alterations to the projected NPP response due to modified phytoplankton nutrient supply in the tropical Pacific has been shown to increase uncertainty in projected impacts of climate change on total consumer biomass (Tagliabue et al., 2020). In another study, accounting for greater complexity in phytoplankton physiology was found to enhance the amplified climate driven responses of higher trophic levels (Kwiatkowski et al., 2019a). Ultimately, ecosystem services depend on changes in absolute NPP rather than the percentage changes that are usually presented. To accommodate the

TABLE 1 | Summary of the ocean provinces used in this study and the codes of the 83 Longhurst provinces (Longhurst, 2007) used by Vichi et al. (2011), with the numbers corresponding to their Table 1 (Vichi et al., 2011).

Province	Long name	Aggregated "Longhurst" provinces
ARC	Arctic	1–3
ASP	Atlantic sub-polar	4, 11, 15
NAS	North Atlantic sub-tropical gyre	5–6, 18
EQA	Equatorial Atlantic	7–9, 12, 17
SAS	South Atlantic sub-tropical gyre	10
IND	Indian Ocean	30–34, 70
SAP	Sub-Arctic Pacific	50–51
NPS	North Pacific sub-tropical gyre	53–56, 60
EQP	Equatorial Pacific	61–63
SPS	South Pacific sub-tropical gyre	59, 64
SOC	Southern Ocean	21, 81–83
GLO	Global Ocean	1–83

The Mediterranean (province 16) is omitted from the regional analysis.

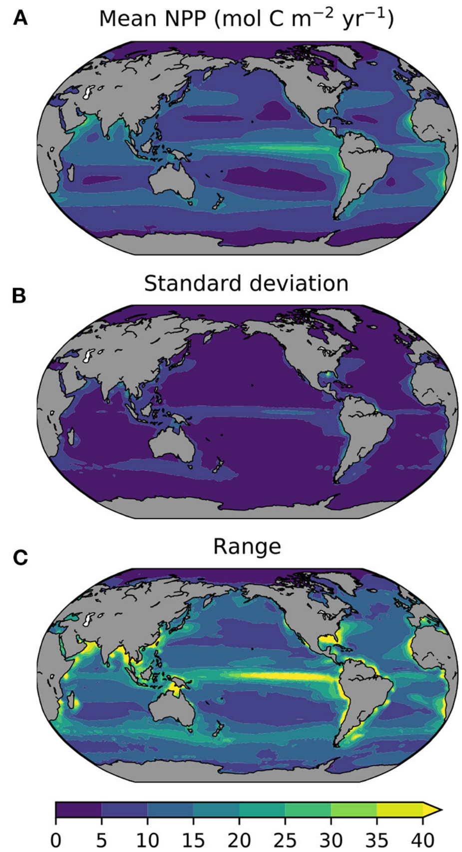

FIGURE 1 | Absolute annual NPP from 16 CMIP6 models for the historical period (1995–2014) presented as **(A)** multi-model mean, **(B)** 1σ, and **(C)** the full inter-model range. All plots show annual NPP in mol C m^{-2} yr^{-1}.

divergent absolute mean state conditions of Earth system models, upper trophic level and fisheries projections are currently forced to rely on proportional changes that are appropriately benchmarked for a subset of Earth system models (e.g., Tittensor et al., 2018).

VARIABILITY IN GLOBAL AND REGIONAL NET PRIMARY PRODUCTION PROJECTIONS FROM CMIP5 AND CMIP6 MODELS

In this section, we examine and review NPP changes from 16 CMIP6 models newly assembled for this study (only 13 were used in Kwiatkowski et al., 2020) and 10 CMIP5 models previously assessed (Bopp et al., 2013). This focus on absolute, as well as relative changes from CMIP6 and CMIP5 models at regional scale is important as they underpin the associated ecosystem services and our understanding of a key component of upper ocean carbon turnover. To facilitate this regional analysis, we re-gridded CMIP6 and CMIP5 models onto a uniform 1 × 1 degree horizontal grid as in Kwiatkowski et al. (2020) and used 11 broad biogeographic provinces based on Longhurst (2007) and summarised previously (Vichi et al., 2011) to represent 11 key oceanic regions (**Table 1**). We address the change in depth-integrated annual NPP averaged between 2081–2100 and 1995–2014 in three ways, the relative change (in percentage terms), the absolute change in mol C m^{-2} yr^{-1} and integrated over the province in Tmol C yr^{-1}. A common reference period is applied to both CMIP6 and CMIP5 models. In all cases, we assessed uncertainty across models in two ways, *via* the one standard deviation "likely" range that links to IPCC assessments and the full inter-model range. We generally focus on using the one standard deviation "*likely*" range, but note the importance of the inter-model range in quantifying the total range of uncertainty across all models. In this way, we update the recent assessments based on CMIP5 and CMIP6 as part of the IPCC sixth assessment reports.

Before addressing the projected changes, we focus on the degree of inter-model difference in the levels of historical (1995–2014) NPP. For example, the multi-model mean global annual NPP is 41.4 Pg (or 3.45 Pmol), but the likely and inter-model ranges across the 16 CMIP6 models are 10 and 33 Pg, respectively. Notably, this inter-model range of 33 Pg C, is over an order of magnitude greater than the projected absolute change (see below). Some models show rates of global NPP as high as 56 Pg C in the historical era, while others are as low as 23 Pg C. These inter-model differences in historical levels of NPP are also present at a regional level, with areas of high absolute NPP in multi-model mean terms (**Figure 1A**), also showing an elevated likely uncertainty (**Figure 1B**) and inter-model range (**Figure 1C**). Notably, the magnitudes of both the likely and inter-model range-based uncertainties are of similar order to the multi-model mean NPP, indicating a high level of inter-model difference (note similar scale range for **Figure 1** panels). In general, the degree of inter-model uncertainty in historical levels of NPP has declined between CMIP5 and CMIP6, with the largest declines in uncertainty across models in regions such as the

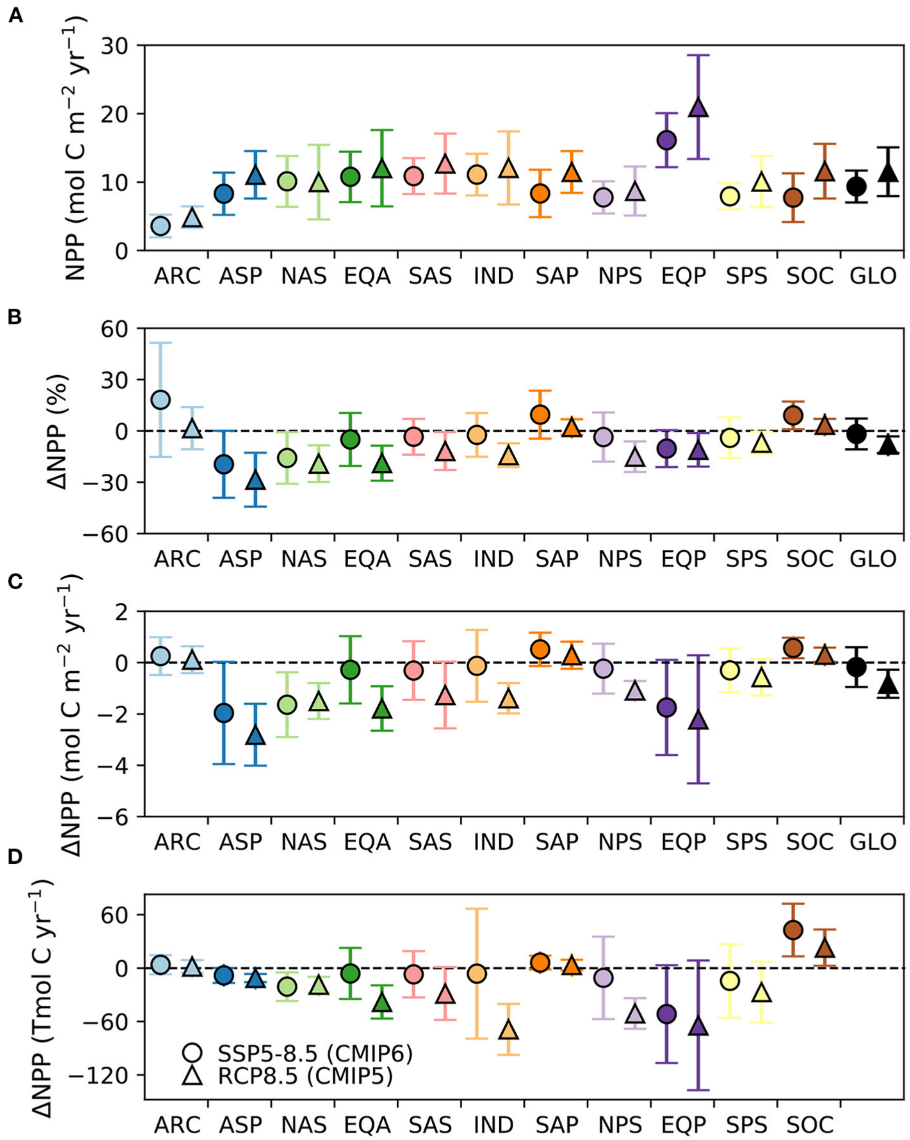

FIGURE 2 | Multi model mean and 1σ uncertainty for 16 CMIP6 and 10 CMIP5 models on a province by province basis for **(A)** NPP levels during the historical era (1995–2014), and projected changes in annual NPP on a **(B)** percent, **(C)** mol C m^{-2} and **(D)** Tmol C basis (all relative to 1995–2014). CMIP6 results are from the 16 models using the SSP5-8.5 scenario, while CMIP5 results are from the 10 models using the RCP8.5 scenario. The global results are not included on panel d as they distort the scale.

tropical Atlantic, Pacific and Indian Oceans (**Figure 2A**). Despite their reduction since CMIP5, these inter-model differences in ocean NPP remain substantial and reflect major mean state biases across models, as well as indicating a large knowledge gap in the magnitude of this key planetary flux.

Projected Changes in NPP at Global and Regional Scales

Relative Changes

In our analysis extended to 16 CMIP6 models, global NPP is projected to decline by 1.76% under SSP5-85, with a likely uncertainty range of ±8.06% or a total inter-model range of 33.96% across all models (relative to 1995–2014). The multi-model change for CMIP6 is slightly less than the analysis of Kwiatkowski et al. (2020), due to the inclusion of three additional Earth system models. The inter-model range of >30% around this small average change highlights the wide array of responses in projections at the global scale and the lack of confidence in even the sign of change for CMIP6. Globally, the multi-model mean change in NPP as declined from around −8.06% for CMIP5 to −1.76% for CMIP6, due to the much larger inter-model spread (**Table 2**). The slight change in the global percentage

TABLE 2 | Statistics (as percentage changes) for the multi-model mean, 1σ "likely" range and the full inter-model range across different regions and globally for 16 CMIP6 and 10 CMIP5 models.

Province	CMIP6	CMIP5	CMIP6	CMIP5	CMIP6	CMIP5
	Multi-model mean ΔNPP (%)		Multi-model 1σ ΔNPP (%)		Inter-model range ΔNPP (%)	
ARC	18.20	1.59	33.34	12.31	125.90	41.19
ASP	−19.50	−28.49	19.59	15.76	71.61	47.98
NAS	−15.80	−19.07	15.12	10.72	58.77	37.34
EQA	−4.97	−18.83	15.52	10.24	53.70	33.79
SAS	−3.43	−11.72	10.44	11.07	28.94	34.53
IND	−2.33	−14.11	12.76	6.95	48.44	23.72
SAP	9.52	2.27	14.04	4.57	45.35	14.60
NPS	−3.60	−15.06	14.39	8.99	50.48	32.59
EQP	−10.33	−11.03	10.81	9.82	36.87	28.85
SPS	−4.04	−6.86	11.93	6.41	43.25	24.45
SOC	9.06	3.62	8.11	3.44	31.61	10.56
GLO	−1.76	−8.06	9.01	4.83	33.96	14.12

Here annual NPP changes are presented in percentage change. Colouring for regions are white centred using all CMIP5 and CMIP6 outputs across the 10–90 percentiles for each metric. CMIP6 and CMIP5 changes are from the SSP5-8.5 and RCP8.5 scenarios, respectively, and from 2081–2100 relative to 1995–2014 for comparison.

change for CMIP5, compared to Bopp et al. (2013), arises due to our use of a later reference period for CMIP5 of 1995–2014.

Except for a few regions, the uncertainty in percentage changes in NPP, measured either by the likely or inter-model range, is broadly similar across regions (**Figure 3**, **Table 2**). Exceptions to this are the Arctic, which shows the highest likely and inter-model range, and the North Atlantic sub-polar gyre (**Figure 3**, **Table 2**). The Southern Ocean and, to a lesser extent, the South Atlantic sub-tropical Gyre, display uncertainties (both in terms of likely and inter-model ranges) that are slightly smaller than the global average. Compared to CMIP5, uncertainties in the projected percentage changes in NPP at regional scale have increased by up to 3-fold under CMIP6, most notably in the sub-Arctic Pacific, Arctic and Southern Ocean, with the exception of the South Atlantic Sub-Tropical Gyre where uncertainty has declined since CMIP5 (**Table 2**, **Figure 2B**).

Absolute Changes

The projected changes in absolute NPP under the high emissions SSP5-8.5 scenario also show inter-model differences that are significant. At the global scale, the projected annual NPP changes are −0.17 mol C m^{-2} (or −63.34 Tmol C) and are associated with likely and inter-model ranges of 0.77 and 2.72 mol C m^{-2} (or 286.40 and 1010.00 Tmol C), respectively (**Tables 3**, **4**). Similar to percentage changes, we find uncertainties in the global projections that are very large, relative to the multi-model mean change. Importantly, the degree of uncertainty in projections of global NPP in absolute terms has grown from CMIP5 to CMIP6 by around 1.5-fold (**Tables 3**, **4**). However, regionally, even stronger levels of uncertainty in absolute NPP changes emerge from the CMIP6 models (**Figure 4**). Regions that hosted high NPP in the historical period (**Figure 1A**) show the largest likely uncertainty and inter-model range for their projected NPP changes in absolute terms (**Figures 4B,C**). The inter-model range in the projected NPP change is amplified 5-fold, relative to the multi-model mean change (note the 5-fold increase in scale for **Figure 4C**, as compared to **Figure 4A**), while the inter-model range was of similar order to the multi-model mean in the historical era (similar range for **Figures 1A,C**).

Turning to the absolute changes at the regional scale in more detail, we see that some key regions emerge as contributing substantially to the uncertainty in projected absolute changes on a per square metre basis (**Figure 5**). Relative to the global uncertainty, the North Atlantic sub-polar Gyre, Equatorial Pacific and the tropical Indian Ocean display the largest uncertainty in the mol C m^{-2} changes in annual NPP for both the likely and inter-model range (**Table 3**, **Figure 2C**). In a regionally integrated sense, the area of the different regions is important and the Indian Ocean, Equatorial Pacific, northern and southern sub-tropical Pacific regions emerge as hotspots of uncertainty (**Table 4**, **Figure 2D**). Indeed, around two-thirds (67%) of the overall inter-model range in the total NPP change across models in a total integrated sense can be explained simply by the summed contributions from the tropical Pacific and Indian Oceans (**Table 4**). Uncertainties, in terms of both the likely and inter-model range, at the regional scale have also expanded markedly since CMIP5, especially for the Indian Ocean and north Pacific sub-tropical gyre where uncertainty has increased around 2–3-fold (**Tables 3**, **4**, **Figures 2C,D**).

Overall, this indicates that a large proportion of the inter-model uncertainty around global NPP projections in absolute terms resides in the response of the Tropical Pacific, Indian and to a lesser extent the North Atlantic Oceans. Some regions that have strong uncertainties in relative changes, like the Arctic (**Figure 3**), are less significant in absolute changes. For the Tropical Pacific and Indian Oceans, the uncertainty in the projections of absolute NPP change concerns the sign of the change, while for the North Atlantic Ocean it relates to the magnitude of the projected decline, which is common across models (**Figure 5**). Lastly, the increase in uncertainty regarding projected changes since CMIP5, despite less uncertainty in the mean state, is notable (**Figure 2**) and will lower confidence in NPP projections made in SROCC due to the reduced inter-model agreement.

It is particularly notable that despite higher equilibrium climate sensitivity (Zelinka et al., 2020), greater surface warming and enhanced upper ocean nutrient decline in CMIP6 than CMIP5, global NPP decline over the 21st century is less, not more, extensive (Kwiatkowski et al., 2020). The lower magnitude of global NPP decline in CMIP6 is not attributable to an individual region, but rather a result of enhanced NPP increases in the Southern and Arctic Oceans and less extensive declines in multiple other regions, notably the Indian Ocean, the equatorial Atlantic and Pacific, and the sub-tropical gyres in the Pacific and south Atlantic (**Figure 2**). This suggests that the temporal evolution of phytoplankton resource limitation and or grazing pressure due to climate-driven changes in atmosphere/ocean physics and/or ocean biogeochemical cycling

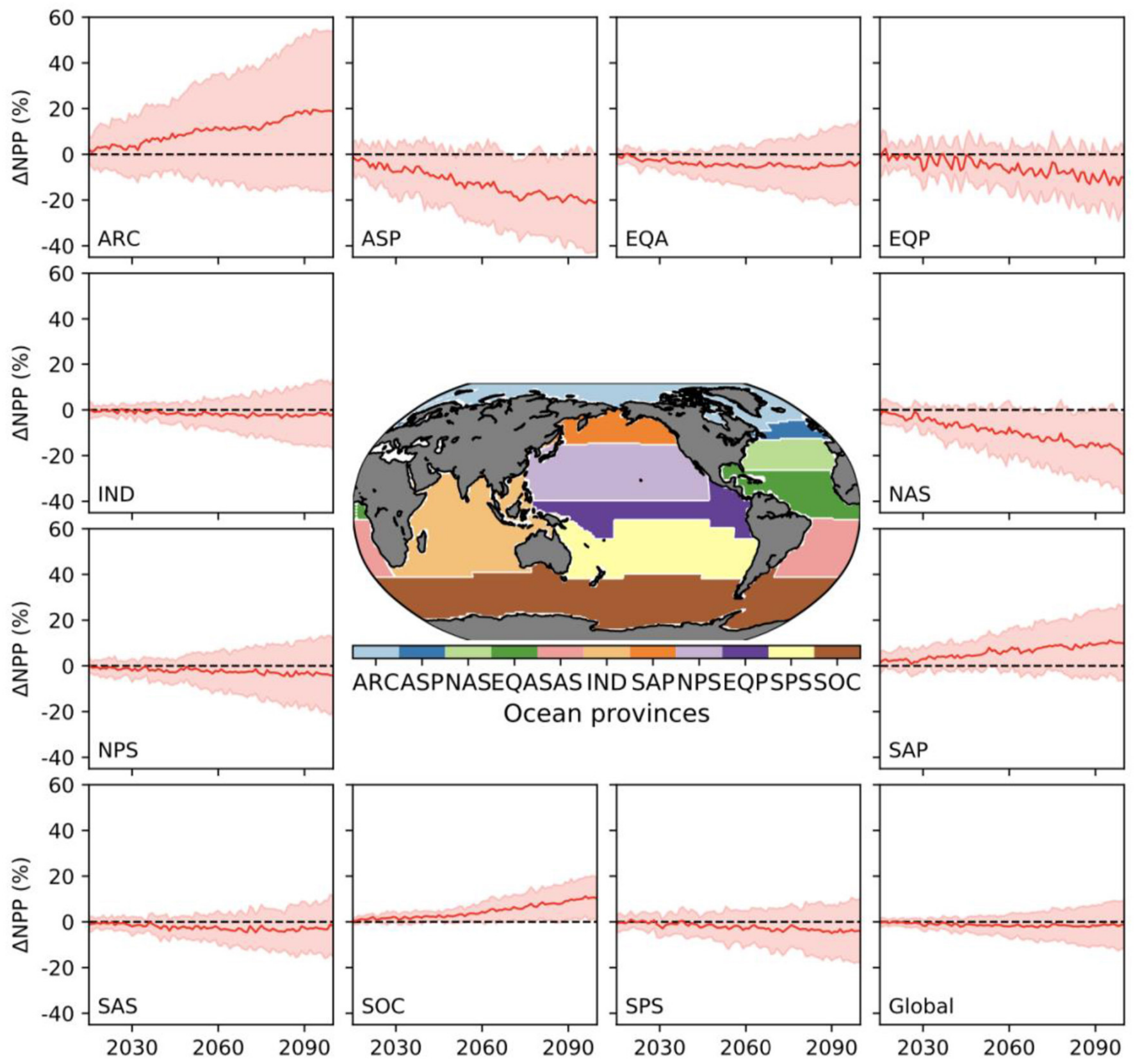

FIGURE 3 | Percentage change in annual NPP (relative to 1995–2014) across different biogeochemical provinces and globally from 16 CMIP6 models. The red line represents the multi-model mean and the shaded area the 1σ "likely" uncertainty range. The map shows the location of the provinces.

may have significantly altered between CMIP5 and CMIP6 in these regions.

Key Mechanisms Underpinning Uncertainty

The dominant conceptual view of how climate change affects NPP in the predominantly nutrient limited low to mid latitude regions is via changes to ocean stratification, which reduce nutrient supplies to the surface ocean (Behrenfeld et al., 2006; Doney, 2006). In contrast, light limited regions at high latitudes benefit from extended growing seasons due to sea ice removal and the maintenance of ocean mixing within the euphotic layer when waters are more stratified. However, this emphasis on stratification is incomplete (e.g., Whitt and Jansen, 2020) and clearly does not account for the variety of physical and biogeochemical mechanisms that produce the inter-model uncertainty in NPP projections for both CMIP5 and CMIP6. In our analysis on the projected changes in NPP in absolute terms, three key regions of the ocean emerged as hotspots of inter-model disagreement: (i) Indian Ocean, (ii) Tropical Pacific (encompassing the equatorial and northern and southern Pacific sub-tropical gyres), and the (iii) North Atlantic (including the sub-polar gyre). Ultimately, the source of inter-model uncertainty in all three regions is 2-fold and linked to the response of (i) atmospheric and ocean circulation changes, and (ii) ocean biogeochemical cycling, including nutrient supply, resource limitation and grazing.

Uncertainties in Atmospheric and Oceanic Circulation Changes

Changes in the tropical Indo-Pacific atmospheric circulation known as the Walker Cell are a good example of how other

TABLE 3 | Statistics (as mol C m^{-2} changes) for the multi-model mean, 1σ "likely" range and the full inter-model range across different regions and globally for 16 CMIP6 and 10 CMIP5 models.

Province	CMIP6	CMIP5	CMIP6	CMIP5	CMIP6	CMIP5
	Multi-model mean ΔNPP (mol C m^{-2})		Multi-model 1σ ΔNPP (mol C m^{-2})		Inter-model range ΔNPP (mol C m^{-2})	
ARC	0.25	0.12	0.74	0.53	2.44	1.89
ASP	−1.96	−2.81	2.00	1.21	7.31	4.46
NAS	−1.64	−1.49	1.26	0.70	4.35	2.20
EQA	−0.28	−1.79	1.31	0.87	4.18	3.04
SAS	−0.31	−1.26	1.14	1.30	3.38	3.87
IND	−0.12	−1.39	1.40	0.59	5.45	2.39
SAP	0.52	0.29	0.65	0.53	2.16	1.75
NPS	−0.23	−1.08	0.97	0.37	2.88	0.98
EQP	−1.75	−2.21	1.86	2.50	6.96	8.37
SPS	−0.30	−0.57	0.86	0.71	2.65	2.97
SOC	0.57	0.31	0.40	0.28	1.35	0.92
GLO	−0.17	−0.82	0.77	0.55	2.73	1.88

Here annual NPP average changes are presented in mol C m^{-2}. Colouring for regions are white centered using all CMIP5 and CMIP6 outputs across the 10–90 percentiles for each metric. CMIP6 and CMIP5 changes are from the SSP5-8.5 and RCP8.5 scenarios, respectively, and from 2081 to 2100 relative to 1995 to 2014 for comparison.

TABLE 4 | Statistics (as Tmol C changes) for the multi-model mean, 1σ 'likely' range and the full inter-model range across different regions and globally for 16 CMIP6 and 10 CMIP5 models.

Province	CMIP6	CMIP5	CMIP6	CMIP5	CMIP6	CMIP5
	Multi-model mean ΔNPP (Tmol C)		Multi-model 1σ ΔNPP (Tmol C)		Inter-model range ΔNPP (Tmol C)	
ARC	3.72	1.64	10.75	7.31	35.61	26.28
ASP	−8.24	−11.17	8.38	4.59	30.65	17.06
NAS	−20.87	−18.76	16.07	8.87	55.33	27.73
EQA	−6.06	−38.04	28.67	18.60	91.33	64.75
SAS	−7.00	−28.62	26.05	29.60	76.98	87.99
IND	−6.34	−68.82	72.93	28.61	284.10	117.00
SAP	6.14	3.37	7.72	6.01	25.56	19.96
NPS	−11.00	−50.96	46.26	17.27	137.00	45.78
EQP	−51.66	−64.38	54.93	72.89	206.10	243.90
SPS	−14.44	−27.04	41.24	33.97	127.60	141.70
SOC	42.69	22.96	29.54	20.45	100.10	67.83
GLO	−63.34	−283.80	286.40	189.30	1010.00	645.40

Here total annual NPP changes are presented in Tmol C. Colouring for regions are white centered using all CMIP5 and CMIP6 outputs across the 10–90 percentiles for each metric. CMIP6 and CMIP5 changes are from the SSP5-8.5 and RCP8.5 scenarios, respectively, and from 2081 to 2100 relative to 1995–2014 for comparison.

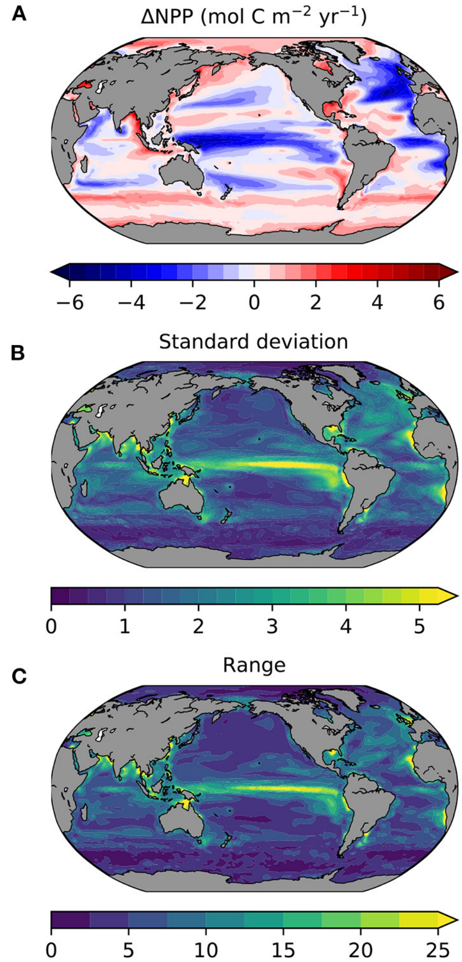

FIGURE 4 | The SSP5-8.5 change in NPP from 16 CMIP6 models by 2081-2100 (relative to 1995–2014) for **(A)** multi-model mean, **(B)** 1σ, and **(C)** the full inter-model range. All plots show annual NPP in mol C m^{-2} yr^{-1}. Note scale change for **(C)**.

factors, such as changing wind patterns, interact with warming to drive NPP change. With easterlies over the equatorial Pacific and Westerlies over the equatorial Indian Ocean, the Indo-Pacific Walker Cell sustains NPP in the eastern Pacific and western Indian Ocean by shallowing the nutricline and making nutrients more available. Any Walker Cell intensification would thus tend to oppose the effect of enhanced water column stability linked to warming, while a Walker Cell slowdown would amplify the NPP decrease in the Pacific by reducing nutrient availability (Matear et al., 2015). Currently, the evolution of the Walker Cell in Earth system models is uncertain, with most projecting a weakening (Cai et al., 2020), but some suggesting an intensification (Plesca et al., 2018). Although observations suggest the Walker Cell has intensified over recent decades, there is a debate regarding whether it can be attributed to a longer term trend that is not consistently reproduced by Earth system models or to internal atmospheric variability (Kociuba and Power, 2015). An improved understanding of how changes in winds, that are poorly constrained, interact with ongoing stratification changes, that are better understood, in driving the overall NPP response in the Indo-Pacific is required.

The western Arabian sea is one of the most productive regions in the entire tropics (Naqvi et al., 2010; Moffett and Landry, 2020) and although changes in stratification contribute to the NPP changes (Roxy et al., 2016), alterations to winds may also be

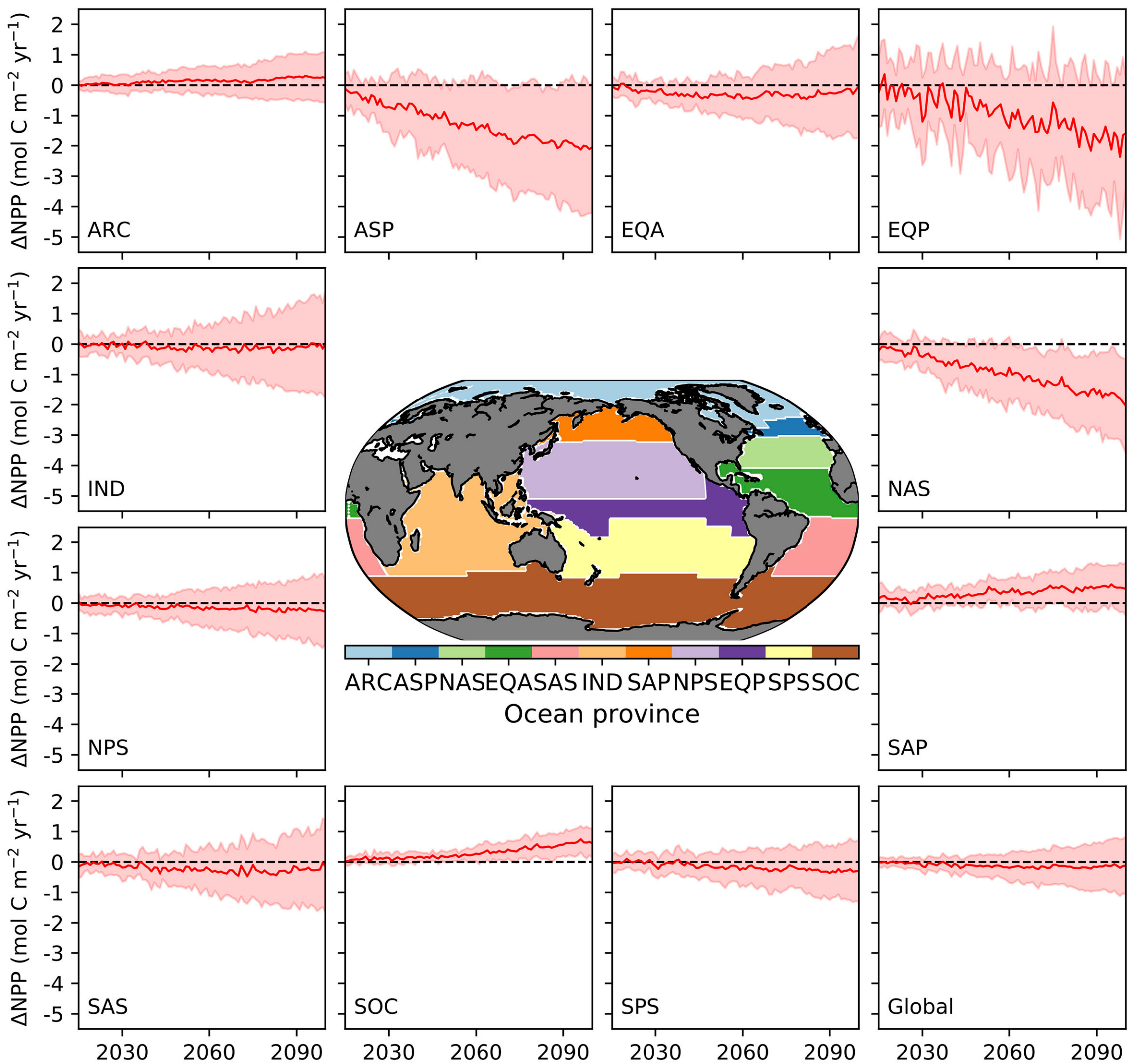

FIGURE 5 | Absolute change in annual NPP (relative to 1995–2014) in mol C m^{-2} across different biogeochemical provinces and globally from 16 CMIP6 models. The red line represents the multi-model mean and the shaded area the 1σ "likely" uncertainty range. The map shows the location of the provinces.

important. Notably, high rates of NPP in the Arabian Sea during summer are supported by the Somalia and Oman upwellings that are driven by alongshore monsoon winds and offshore Ekman pumping, with mesoscale processes also playing an important role in its horizontal extent (e.g., Marra and Barber, 2005). In contrast, during winter, Arabian Sea NPP is regulated by convective mixing in response to winter monsoon evaporative cooling (e.g., Keerthi et al., 2017). Most Earth system models project a decrease in the monsoon winds over the Arabian Sea during both winter (Parvathi et al., 2017) and summer (Sooraj et al., 2014), as well as anomalous easterlies at the equator in response to the Indo-Pacific Walker circulation weakening. These changes will contribute to an NPP decline during both seasons, but are highly variable across models (as discussed above). Ultimately, significant biases and uncertainties in the representation of the Asian monsoon and its oceanic response in Earth system models hamper the reliability of NPP projections in the Indian Ocean (Singh et al., 2019). As fine-scale processes, such as mesoscale eddies, are very important in this region, improvements to model resolution during CMIP6 (Eyring et al., 2016) have the potential to improve the reliability of future NPP projections.

Changes in coastal upwelling in the Eastern Boundary Upwelling Systems (EBUS) due to shifts in wind patterns will also be important in shaping the response of NPP and the coupling to ecosystem services. Early studies suggested strengthening winds due to climate change would lead to consistent increases in upwelling in EBUS, with implications for the response of NPP (Bakun, 1990). Current Earth system models, and some observations, tend to project reduced winds and upwelling intensity for low latitudes and enhancements at higher latitudes under a high emissions scenario during CMIP5 (García-Reyes et al., 2015; Rykaczewski et al., 2015). However, the interaction of changing winds with coastal warming can produce complex responses in upwelling at local scales, with an important role for mesoscale dynamics not resolved in Earth system models (García-Reyes et al., 2015; Wang et al., 2015; Renault et al., 2016; Xiu et al., 2018).

Interannual changes in NPP in the North Atlantic have been shown to be poorly correlated with parallel modifications to stratification (Lozier et al., 2011; Dave and Lozier, 2013), highlighting the role of other physical mechanisms. The Gulf Stream plays a critical role in nutrient supply in the North Atlantic via the nutrient stream (Pelegrí et al., 1996). This means that projected changes to nutrient concentrations in the upper ocean from Earth system models result from the combination of the nutrient stream and its sensitivity to the Atlantic meridional overturning circulation (AMOC), which interact with the effect of changes to ocean stratification (Tagklis et al., 2020; Whitt and Jansen, 2020; Couespel et al., 2021). Differences in the mean state representation of the nutrient stream and the extent of AMOC slowdown across CMIP5 (Cheng et al., 2013) and CMIP6 (Weijer et al., 2020) are therefore major contributors to the magnitude and uncertainty of projected NPP declines in the region (Tagklis et al., 2020; Whitt and Jansen, 2020).

Uncertainties in the Representation of Biogeochemical Processes

In the tropical Pacific, in particular, part of the inter-model uncertainty is driven by the relative role played by iron and nitrogen limitation among models. A set of mechanistic experiments within a single model demonstrated that minor parameter adjustments that altered the strength of iron limitation and its replacement by nitrogen limitation in the tropical Pacific led to large changes in both the magnitude and sign of changes to regional NPP (Tagliabue et al., 2020). It is highly likely that there is divergence in the representation of nutrient limitation regimes across CMIP5 and CMIP6 models, in part due to the paucity of observational constraints (Moore et al., 2013; Ustick et al., 2021). Indeed, it was notable that the emergent constraint that assessed NPP projections from CMIP5 using the co-variation of sea surface temperature and NPP anomalies (Kwiatkowski et al., 2017), was not able to strongly discriminate among different iron-nitrogen limitation scenarios (Tagliabue et al., 2020). New understanding to inform on Earth system model parameterisations and performance are clearly needed for this key aspect of model structural uncertainty.

Poor knowledge of nutrient limitation regimes is not restricted to the tropical Pacific, but also the Indian Ocean, with local physical processes also being important. Our knowledge of Indian Ocean biogeochemical cycling and nutrient limitation regimes remains limited but has been growing in recent years (Grand et al., 2015; Chinni et al., 2019; Twining et al., 2019). For instance, iron limitation and zooplankton grazing have been suggested to modulate NPP in the western Arabian Sea (Moffett et al., 2015; Moffett and Landry, 2020), which highlights the need for a deeper understanding of how grazing processes and iron limitation interact in this highly productive region. Additionally, if iron limitation is important, then climate-driven changes in dust fluxes will also be a key component of projected changes in Earth system models. Of course, these biogeochemical changes also interact with the atmospheric and ocean physics changes mentioned in section Uncertainties in Atmospheric and Oceanic Circulation Changes to control the fate of NPP. As suggested by Moffett and Landry (2020), any strengthening of the monsoonal circulation should exacerbate iron limitation, leading to an eastward shift in the utilization of upwelled nutrients, while a weakening of the monsoonal circulation, as projected by most climate models (Singh et al., 2019), will probably reduce the importance of iron limitation. It is likely that there is ultimately a strong mosaic of nutrient limitation regimes in the region, linked to both the underlying physical finescales, but also the role of dust, river and margin supplies that needs to be assessed further in Earth system models.

Relevant across all the tropical regions showing high uncertainty is the contribution played by nitrogen fixation by diazotrophs in modulating the NPP response to climate change. As a new source of nitrogen to the upper ocean, nitrogen fixation may counterbalance changes in nutrient supply due to changing ocean stratification, but only if sufficient iron and phosphorus is present (Zehr and Capone, 2020). Projected changes to nitrogen fixation are highly variable among models at regional scale and because modifications to nitrogen fixation due to climate change emerge rapidly, they can contribute to the changes in NPP in nitrogen limited regions that establish themselves outside of internal variability later in the century (Wrightson and Tagliabue, 2020). CMIP5 and CMIP6 models take different approaches to representing diazotrophy, which introduces an additional source of uncertainty in the evolution of the upper ocean fixed nitrogen supply.

In the Arctic, NPP projections are particularly uncertain in areas of present-day sea ice, where initial nitrate concentrations and the rate of sea ice retreat determine how reduced light limitation and enhanced nutrient limitation interact to determine the future NPP response in a given model (Popova et al., 2012; Vancoppenolle et al., 2013). In the Arctic also, a recent study has also shown that nutrient input from the land, through riverine fluxes and coastal erosion, may sustain around one third of contemporary Arctic Ocean NPP (Terhaar et al., 2021). This suggests that changes in nutrient input may be a key factor affecting future Arctic NPP, in accord with recent satellite derived trends (Lewis et al., 2020).

More broadly, additional sensitivities around the linkage between phytoplankton and zooplankton, as well as upper ocean nutrient cycling will contribute to inter-model uncertainties. The ability of Earth system models to resolve variations in the nutrient contents of phytoplankton plays a role in their responses to changing nutrient supply (Kwiatkowski et al., 2018), with implications for zooplankton grazers (Kwiatkowski et al., 2019a). Differences in the changes in grazing rates across Earth system models contributes additional uncertainty to NPP projections, due to their impact on overall phytoplankton biomass (Laufkötter et al., 2015). Moreover, adjustments to zooplankton grazing efficiencies due to climate change induced changes in food quality can lead to alterations to and feedbacks on upper ocean recycling, especially for essential micronutrients

(Richon et al., 2020; Richon and Tagliabue, 2021). Whether or not temperature sensitivities are applied to nutrient recycling pathways in Earth system models may also affect the temporal evolution of NPP (Taucher and Oschlies, 2011). Finally, explicit representation of the "end to end" coupling between phytoplankton and upper trophic levels is not yet widespread, but may also be important in shaping the response of NPP (Lefort et al., 2015; Aumont et al., 2018).

IMPLICATIONS FOR OCEAN BIODIVERSITY AND ECOSYSTEM SERVICES

The large uncertainties regarding the projected changes in NPP in extent and sign, as well as their increase from CMIP5 we report here, have important implications for fisheries management. This is particularly the case because uncertainties increase at regional scales and are accentuated in regions where fisheries contribute substantially to food provision (e.g., the Tropical Pacific and Indian Ocean) (Golden et al., 2016; Bindoff et al., 2019). Temperature has been found to be a principal driver of changes in the biogeography of marine fish and invertebrates, while NPP interacts with temperature to affect the productivity of the populations (Brander, 2007; Blanchard et al., 2012; Cheung et al., 2019). Thus, production of marine species that expand poleward in response to warming may be limited by the decrease in NPP in their future range, resulting in a decline in total fish production as the low latitude edge of the distribution range retract poleward (Jones and Cheung, 2015; Villarino et al., 2015; Cheung et al., 2019). In contrast, increase in NPP may increase the rate of range expansion. Simultaneously, changes in NPP, as the basic energy production, will also result in cascading effects through the food chains from zooplankton to fish and top predators such as seabirds and marine mammals (Bindoff et al., 2019; Tagliabue et al., 2020; du Pontavice et al., 2021). In relation to this, an added dimension of uncertainties in NPP projections that are important for understanding the effects on marine foodwebs is the nutritional quality associated with NPP (e.g., size, structure and nutrient content). These changes will impact the availability of seafood from fisheries, both globally and regionally. Because the uncertainty range of NPP is high, and has increased since the CMIP5 results available for the SROCC, there are rising challenges for the long-term planning of fisheries development. These include the need for strategies that are more adaptive to an increasingly uncertain future. Similar considerations would also hold for other ecosystem services, based on the fact that coastal and shelf seas are providing up to 50% of global ocean primary production, thereby supporting a variety of rich habitats, such as wetlands, mangroves, kelp forests, and seagrass meadows that support biodiversity and provide food, coastal protection, carbon sequestration, and other services (Bindoff et al., 2019).

In terms of the mitigation of climate change, the uncertainties in the reduction of the biological carbon pump translate to differences spanning orders of magnitude in abatement and social costs, particularly given the increased uncertainty with respect to previous estimates from CMIP5 and related cost estimates (Barange et al., 2017). For instance, in the Barange et al. (2017) study, the projected decrease in carbon sequestration in the North Atlantic of 27–41% was estimated to represent a loss of 170–3,000 billion USD in abatement (mitigation) costs and 23–401 billion USD in social costs. While NPP is a key first step, an array of additional processes lead to the eventual change in the biological carbon pump. For instance, parallel changes in the plankton community composition, sinking rates and the remineralisation of particles will decouple changes in the biological carbon pump from NPP (De La Rocha and Passow, 2014; Sanders et al., 2014). Nevertheless, due to their connectivity, increased uncertainty in NPP will certainly not increase confidence in projections of the biological carbon pump. However, it is possible that additional constraints, e.g., from ocean interior nutrient and oxygen levels, can better constrain the biological carbon pump and thus bolster confidence in assessments.

FUTURE EFFORTS TO REDUCE UNCERTAINTY IN NPP PROJECTIONS

A major question is whether the community will be able to constrain ocean NPP better in the future, so model skill and inter-model agreement for NPP looks more like it does for nitrate and phosphate. The major issue for NPP is the high degree of variability across different spatial and temporal scales (compared to whole ocean nutrient reservoirs) and large divergence in contemporary direct constraints, which rely on different remote sensing algorithms applied across different satellite sensors (Sathyendranath et al., 2020). Unlike the biological carbon pump, there is only limited ability to constrain the magnitude and distribution of NPP indirectly from nutrient and oxygen distributions. This suggests that new approaches, such as using measurements like carbon isotopes or genomics, allied to improved observational coverage from *bioArgo*, are needed to provide the improved picture of both the magnitude and variability of upper ocean NPP that is required to drive improved models.

Associated with the challenge of constraining the magnitude of present-day NPP is uncertain knowledge of the drivers of change and how inter-model differences in atmospheric and ocean physics cascade through ocean biogeochemical cycles to drive resource limitation of NPP. Knowledge is growing regarding the role of multiple concurrent factors regulating microbial activity (Saito et al., 2014; Browning et al., 2017; Caputi et al., 2019; Ustick et al., 2021). But how the emergence of these multiple drivers are affected by the climate responses of key physical components, like winds, in the Indo-Pacific region in particular, is a major knowledge gap. Future efforts that conduct additional model experiments aimed at addressing the role of key mechanisms (e.g., Tagliabue et al., 2020), rather than the ensemble of differences (across land, atmosphere, and ocean components) between different Earth system models, are urgently needed to advance this area.

Largely absent from discussions around changing NPP in response to changing nutrient supply is the role of modification to external nutrient inputs to the ocean. While there have been some efforts including dynamic coupling with aerosol nutrient supply at the sea surface (e.g., Yool et al., 2021), future work that accounts for additional external inputs may introduce additional uncertainty in NPP changes. For instance, the inclusion of dynamic Greenland and Antarctic ice sheets in Earth system models is rapidly becoming computationally permissible and the associated changes to freshwater fluxes due to climate change have the potential to influence projections of NPP *via* both physical and biogeochemical mechanisms (Hopwood et al., 2020). These include changes to stratification and associated consequences for nutrient supply (Vizcaíno et al., 2008), but also direct supply of key nutrients, such as iron (Laufkötter et al., 2018; Person et al., 2019) and low latitude impacts via atmospheric teleconnections (Kwiatkowski et al., 2019b). Earth system models are also moving toward dynamic representations of land-ocean riverine nutrient fluxes (Séférian et al., 2020), which have a strong sensitivity to anthropogenic activity and management choices (Seitzinger et al., 2010). The choice of which riverine nutrients to resolve and how these fluxes are coupled to land surface models has the potential to substantially alter projections of NPP in coastal regions (e.g., Terhaar et al., 2019).

At present, each CMIP Earth system model typically differs from its parent generation in terms of both physical and biogeochemical ocean models, not to mention differences in other components of the coupled Earth system (Bopp et al., 2013; Kwiatkowski et al., 2020; Séférian et al., 2020). This presents an inherent traceability issue when trying to identify what determines differences in NPP projections both within and across CMIP generations. One approach that has been previously adopted is to couple a suite of ocean biogeochemical models with the same physical ocean model (Friedrichs et al., 2007; Kriest et al., 2010; Popova et al., 2012; Kwiatkowski et al., 2014) in order to isolate the role of biogeochemistry. Although often practically difficult, this can be a valuable exercise, identifying the limitations of specific model parameterisations. Another area where improvements could be made is in the documentation of Earth system models, specifically, with respect to model spin-up procedures (e.g., Séférian et al., 2016) and validation. Currently each ocean biogeochemical modelling group makes independent and undocumented validation decisions. Knowledge of what observational constraints have been used to constrain ocean biogeochemical models would critically improve our understanding of the uncertainty in NPP projections. In this respect, much could be learned from the terrestrial biogeochemistry community (e.g., Spafford and MacDougall, 2021).

CONCLUDING REMARKS

Overall, the increase in model uncertainty between CMIP5 and CMIP6 for projections of NPP dramatically exceeds that of other ecosystem drivers such as sea surface temperature (Kwiatkowski et al., 2020). However, one might also consider whether the analysis we have made here constitutes a realistic assessment as to the magnitude of the uncertainty in the projected changes to ocean NPP. Despite progress in recent years, the ocean biology components of Earth system models remain very simple, relative to the complex picture of the structure and function of the ocean microbiome that is emerging from observations (Sunagawa et al., 2015). Future work that explores the consequences of new processes in Earth system models, including (but not limited to) microbial co-limitation, acclimation or adaptation, mixotrophy, and predator-prey dynamics will raise challenges for constraining model uncertainty (especially at regional scales) and model democratisation (Sanderson et al., 2015). This will be especially true for efforts using ocean biology changes from Earth system models to project upper trophic levels and provide assessments of fish stock changes. More broadly speaking, making the link between upper trophic levels and NPP in the context of a changing climate would benefit from moving beyond the single forcing of anthropogenic climate change toward a holistic framework of impacts on ocean ecosystems, ranging from pollutants (e.g., mercury, persistent organic pollutants, plastics) to disease (e.g., *Vibrio*) to overfishing.

DATA AVAILABILITY STATEMENT

Publicly available datasets were analyzed in this study. This data can be found at: https://esgf.llnl.gov/.

AUTHOR CONTRIBUTIONS

The study was designed and led by AT. Analysis of CMIP5 and CMIP6 models was performed by AT and LK. AT led the writing of the manuscript with contributions from LK, LB, MB, WC, ML, and JV. All authors contributed to the article and approved the submitted version.

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Climate Change Impacts on Polar Marine Ecosystems: Toward Robust Approaches for Managing Risks and Uncertainties

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The Polar Regions chapter of the Intergovernmental Panel on Climate Change's Special Report on the Ocean and Cryosphere in a Changing Climate (SROCC) provides a comprehensive assessment of climate change impacts on polar marine ecosystems and associated consequences for humans. It also includes identification of confidence for major findings based on agreement across studies and weight of evidence. Sources of uncertainty, from the extent of available datasets, to resolution of projection models, to the complexity and understanding of underlying social-ecological linkages and dynamics, can influence confidence. Here we, marine ecosystem scientists all having experience as lead authors of IPCC reports, examine the evolution of confidence in observed and projected climate-linked changes in polar ecosystems since SROCC. Further synthesis of literature on polar marine ecosystems has been undertaken, especially within IPCC's Sixth Assessment Report (AR6) Working Group II; for the Southern Ocean also the Marine Ecosystem Assessment for the Southern Ocean (MEASO). These publications incorporate new scientific findings that address some of the knowledge gaps identified in SROCC. While knowledge gaps have been narrowed, we still find that polar region assessments reflect pronounced geographical skewness in knowledge regarding the responses of marine life to changing climate and associated literature. There is also an imbalance in scientific focus; especially research in Antarctica is dominated by physical oceanography and cryosphere science with highly fragmented approaches and only short-term funding to ecology. There are clear indications that the scientific community has made substantial progress in its ability to project ecosystem responses to future climate change through the development of coupled biophysical models of the region facilitated by increased computer power allowing for improved resolution in space and time. Lastly, we point forward—providing recommendations for future advances for IPCC assessments.

Keywords: Arctic, Antarctic, Southern Ocean, polar, marine, ecosystem, SROCC, IPCC

INTRODUCTION

The Polar Regions chapter of the Special Report on the Ocean and Cryosphere in a Changing Climate (SROCC) by the Intergovernmental Panel on Climate Change (IPCC) provides a broad overview and assessment of climate change impacts on polar marine ecosystems and associated consequences for humans (Meredith et al., 2019). The report provides a comprehensive assessment of the current state of scientific findings regarding climate change related to key concepts of risk, adaptation, resilience and transformation (Garschagen et al., 2019). The report uses calibrated language modified from previous approaches (Mastrandrea et al., 2010; Mach et al., 2017) to depict the uncertainty around conclusions in a clearly defined and consistent manner. This calibrated language approach allows for qualitative and quantitative assessments of the confidence of the scientific community in a key finding. Projections of future ocean conditions are based on scenarios of future climate conditions based on a common suite of pathways (representative concentration pathways and shared socioeconomic pathways, Abram et al., 2019). Following IPCC procedure, the report underwent a large and rigorous peer review prior to publication, conducted by the global science community and national science reviews; author teams must provide responses to all reviews. As such, the report represents the global scientific community's best attempt to report the state of the science on climate change with respect to key findings and the confidence there-in at the time the report was written. These special reports form part of a regular cycle by the IPCC, updating its assessments every 7 years, thereby tracking changes in knowledge and understanding over time.

The IPCC is currently preparing its Sixth Assessment Report (AR6) and literature on polar marine ecosystems published since SROCC is being assessed in Working Group II (impacts, adaptation and vulnerabilities). This comprehensive work – in addition to published papers that contribute advances in scientific understanding (e.g., the Marine Ecosystem Assessment for the Southern Ocean, MEASO, 2020)–provides increased evidence for widespread impacts of climate change on polar regions, and new insights on approaches for adapting to imminent climate impacts. A remaining primary focus is to evaluate the feasibility and effectiveness of approaches to reduce negative consequences and retain resilience, termed "adaptations." Conclusions by Working Group 1 for AR6 (The physical science basis; IPCC, 2021) are consistent with earlier assessment reports that climate change is altering polar environments at unprecedented rates. Further, these regions cannot be fully shielded from the effects of climate change through adaptation alone; adaption effectiveness is substantially enhanced by global carbon mitigation.

Despite advancements in understanding, knowledge gaps remain that influence assessment confidence around the magnitude, timing, and scale of impacts and adaptation effectiveness. To identify these knowledge gaps, and measures to address them, we have assembled a team of all four lead authors of the marine ecosystem part of the Polar Regions chapter in SROCC and three AR6 WGII lead authors on polar marine ecosystems. Two of our team were also lead authors in the Polar Regions Chapter of AR5 (Larsen et al., 2014; Meredith et al., 2019). Our paper considers how scientific uncertainty has evolved since AR5 and SROCC and what the implications are for decision-making. Further, we consider the implications of current expectations and orientations of the IPCC for ongoing capacity to assess future climate change impacts for social-ecological systems in the polar regions and options for responding to these impacts. Finally, we put forth recommendations for future advances for IPCC assessments.

EVALUATION OF ADVANCES IN POLAR MARINE ECOSYSTEM KNOWLEDGE SINCE SROCC

Data Gaps and Skewness in Data Coverage

SROCC and AR6 are assessments based upon recent scientific advancements derived from monitoring, process studies, retrospective studies, laboratory investigations, new technologies and modeling studies. In polar regions, improved observational coverage, particularly in regions with low coverage today, will greatly improve the foundation for future assessments. For instance, earth observation systems generally are well-coordinated in large programs (like USA's NASA EOS and EU's Copernicus CMEMS). Therefore, most of the literature on production in the Arctic as a whole region is based on satellite data (Arrigo and van Dijken, 2011, 2015; Kahru et al., 2016) or bioclimatic modeling (Cheung et al., 2009; Fernandes et al., 2020). Scenarios for lower trophic levels from the Regional Ocean Modeling System Nutrient-Phytoplankton-Zooplankton Model (Bering10K ROMSNPZ; Hermann et al., 2016, 2019; Pilcher et al., 2019; Kearney et al., 2020) now also provide alternative perspectives that were still nascent during the SROCC process.

For *in situ* measurements, which are essential for understanding ecosystems at levels above primary production, the situation is quite different (**Figure 1**). Formation of a comprehensive, calibrated, multinational ecosystem monitoring program is, for both the Antarctic and Arctic, challenging given the high cost of sampling in distant regions of the planet. In the interim, nations are striving to improve coordination of existing measurement programs and access to data, such as through the Southern Ocean Observing System (Newman et al., 2019). However, classic *in situ* observations, like research cruises, are to a large degree still organized at a national or smaller regional level (but see, e.g., Eriksen et al., 2018 for joint Norwegian-Russian long term monitoring of the Barents Sea).

Still, considerable progress is being made in some fields, mostly related to ocean physics and biogeochemistry. This includes technological developments and an increasing deployment of autonomous observation platforms that utilize the most developed of these technologies, like Argo profiling floats, drifters, gliders, fixed moorings, vessels of opportunity, and Ice-Tethered Platforms. This is expected to gradually reduce the dependence on ship-based surveys for the collection of physical and biogeochemical variables. Marine mammal or fish species at or near the ocean surface (like tuna and mackerel) and

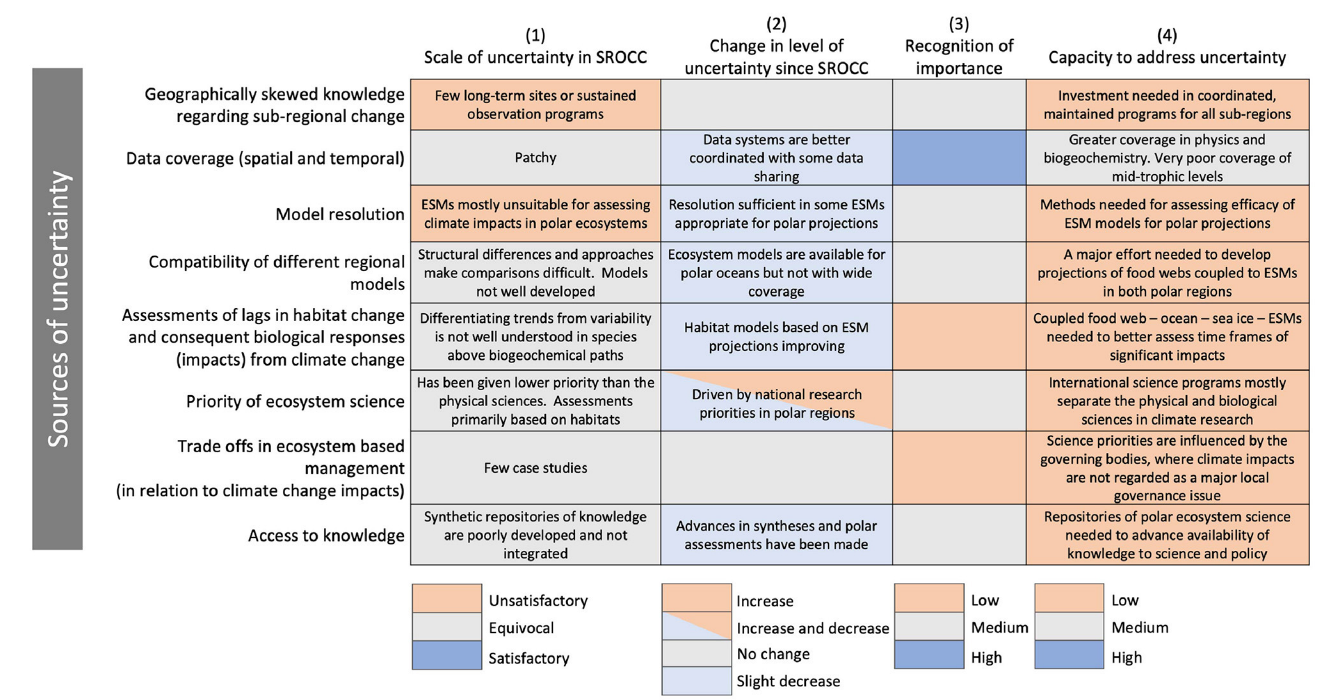

FIGURE 1 | Summary of capacity to advise on, and prognoses for, the impacts of climate change on ecosystems of the polar regions. For different sources of uncertainty, qualitative judgement is provided on (1) the scale of uncertainty in SROCC (blue, satisfactory; gray, equivocal; red, unsatisfactory); (2) change in the level of uncertainty since SROCC (blue, decrease; gray, no change; red, increase); (3) level of recognition of the importance of each source of uncertainty by research and policy communities, i.e., recognition that each source needs to be addressed (blue, high; gray, medium; red, low); and (4) existing capacity to address each source of uncertainty in assessment and decision making (blue, high; gray, medium; red, low). Cells indicate important scientific issues causing uncertainty.

anadromous fish species that return to spawn in rivers and lakes can to some degree be monitored by means of earth observation sensors on satellites or aerial surveillance technologies like Light Detection and Ranging (LIDAR). There are also some examples of *in situ* unmanned measurement of biological variables, for instance autonomous samplers that show promise for remote measurement of the abundance and spatial distribution of pelagic fish (Verfuss et al., 2019). Also, increased activity in high latitude regions, with e.g., polewards displacement of fisheries, and increased tourism, increase options for sampling from ships of opportunity (e.g., Escobar-Flores et al., 2020). However, in the forseable future most biological measurements will still depend on research vessel surveys.

In the Arctic, groups like SAON (Sustaining Arctic Observing Networks) and shorter-term projects like INTAROS (Integrated Arctic Observation System) are making progress toward long-term Arctic-wide observing activities that provide free, open, and timely access to high-quality data, also on ecological variables [see for instance Datasets–INTAROS Data Catalog (nersc.no) and https://doi.org/10.5194/essd-13-1361-2021]. For Antarctica, the Southern Ocean Observing System (SOOS) was implemented after the OceanObs'09 conference and has been instrumental for ocean and cryosphere observations in the Southern Ocean region from a climate driven perspective (Newman et al., 2019).

The science supporting SROCC on climate impacts on marine polar ecosystems reflects regional investment in monitoring and research and is strikingly skewed within and between regions (**Figure 1**). SROCC shows that more scientific results are available for the Arctic than the Antarctic. In the Arctic, there is far more published material on marine life in the eastern Bering and Barents Seas than for the remaining Arctic shelf seas. For instance, marine life in the Kara Sea is mentioned once, the Laptev Sea not at all. For the Southern Ocean, the South Atlantic and west Antarctic Peninsula region are by far the most reported on (see Meredith et al., 2019). International partnerships such as Marine Ecosystem Assessment for the Southern Ocean (MEASO) and the Arctic Monitoring and Assessment Programme (AMAP) are designed to coordinate the collection and analysis of marine life in polar regions. Such coordinated research is providing much needed observations that provide insights into the structure and function of polar ecosystems, thus informing IPCC assessments and providing help to address this regional imbalance in research.

In IPCC's 5th assessment report (AR5; Hollowed and Sundby, 2014; Larsen et al., 2014) and SROCC (Meredith et al., 2019) authors pointed to spatial heterogeneity in ecosystem responses to projected climate change. These occur in response to different physical and biogeochemical changes in the shelf seas and the Arctic Ocean or sea ice habitats in the Southern Ocean, and only strengthens the need for more knowledge on the less understood seas. However, for both the Antarctic and Arctic, closer examination of literature in languages other than English (e.g., Chinese, Japanese, Korean, Polish, Russian, Spanish and Portuguese) is likely to provide important information. Although IPCC authors are also charged with examining this literature, its accessibility is lower and with limited time available such sources of knowledge may be overlooked. In addition to language barriers, results and dissemination from some long-term monitoring programs may not find their way into primary literature due to outdated methodology or the particular focus/question considered to be of regional/local scope only. Yet, in under-sampled regions, such data and corresponding (non-English) regional reports may narrow knowledge gaps.

There is no quick fix to developing time series in under-sampled regions, retrieval of data from non-archived programs, or for making literature generally available from across the diversity of languages. These issues transcend individual experts that become lead authors in the IPCC. The responsibility for resolving them rests, institutionally, with the international scientific community and would be an important topic for consideration and resolution within the partner institutions of the IPCC and by the International Science Council.

Indigenous, Traditional, and Local Knowledge

A source of knowledge increasingly recognized as important to the work of the IPCC Working Group II, is Indigenous and Traditional Knowledge (ITK). This is because of what ITK offers for understanding climate change impacts on ecology as well as on social and human systems.

Indigenous Knowledge of Arctic marine social and ecological systems is spatially and temporally broad, especially with respect to climate impacts on Arctic social-ecological systems and effectiveness of adaptation responses. For the Arctic, ITK can provide important insights needed for understanding current climate change impacts, efficacy of adaptation measures, and future conditions and risks (Petzold et al., 2020; Van Bavel et al., 2020; Eerkes-Medrano and Huntington, 2021; Hauser et al., 2021). While the IPCC's 5th assessment report identified the need to consider information from multiple knowledge sources – and SROCC made progress in this regard for polar regions– inclusion of ITK has been hindered by inconsistent methods for participation and inclusion, and where included is often general in scope and lacks a robust and nuanced treatment of the inter-complexities of climate change impacts and colonialism (Ford et al., 2016, 2020; Petzold et al., 2020). ITK includes methodologies and peer-review processes that are different in ontology and axiology from mainstream science. Therefore, a more inclusive approach to authorship (i.e., including researchers with ITK as lead authors on IPCC reports), promotion and support of participatory and collaborative input from multiple ITK holders (e.g., as contributing authors in addition to ITK expert Lead Authorships) and broader methodology for knowledge review is needed to support greater understanding and more thorough assessment of climate impacts, risks, and responses in Arctic systems.

Language Barriers

Further, knowledge relevant to local systems and management are often published outside of (mainstream) journals and in languages other than English. Such knowledge thus receives less attention by peers and becomes hard to access in global assessments like SROCC (Muelbert et al., 2021). Smaller locally or regionally run assessments, feeding in to global reporting processes like those by IPCC, would help make suck knowledge more available.

Projections of Ecosystem Responses

Climate change will cause polar marine environments to warm further with cascading effects on sea ice extent and thickness, changing the productivity of species and the relative importance of different energy pathways through food webs (Meredith et al., 2019; Trebilco et al., 2020; IPCC, 2021; Thorson et al., 2021). Knowledge of the direction and magnitude of these changes in marine life has incrementally evolved since AR5. Conclusions from AR5 (Larsen et al., 2014) and SROCC (Meredith et al., 2019) highlight compelling evidence that the cascading effects of shifts in the timing and magnitude of seasonal biomass production could disrupt matched phenologies. The authors of the 2014 AR5 Polar chapter identified that a key knowledge gap was that ecosystem models of ecological and social systems at that time were either lacking or insufficiently validated to project the cascading effects of climate change into the future (Larsen et al., 2014). By the time SROCC was completed, projections from global models of marine life based on dynamic bioclimate envelope models were published, and this was cited in SROCC (Bindoff et al., 2019).

Since SROCC a number of studies have projected the future state of polar ecosystems, using both species distribution, multispecies, and Ecosystem-based models (EBMs, Hansen et al., 2019a; Huckstadt et al., 2020; Reum et al., 2020; Veytia et al., 2020; Rooper et al., 2021; Whitehouse et al., 2021). As ecology has no agreed-upon set of mathematics to dictate system dynamics, there is substantial heterogeneity in the theoretical underpinning of these models, processes considered, parameterizations, spatial extent and taxonomic, spatial and temporal resolution, which have implications for model dynamics and outcomes (Payne et al., 2016; Tittensor et al., 2018). Also, as for the Earth System models of the climate systems, these models are subject to uncertainties that may be structural (relating to model parameterization) or parametric (i.e., imperfect measurements, natural variability and abstractions within the model) (Payne et al., 2016; Tittensor et al., 2018). These uncertainties, and additional scenario-based uncertainties (e.g., uncertainties related to future changes in emission rates, land and sea use practices), point to the benefits of ecosystem model intercomparisons, such as the Fisheries and Marine Ecosystem Model Intercomparison Project (Fish-MIP, Tittensor et al., 2018) or the Alaska Climate Integrated Modeling project (ACLIM; Hollowed et al., 2020). Intercomparisons of different models help identify outcomes that are similar despite variability in model construction, giving confidence in those conclusions (Peck et al., 2018; Tittensor et al., 2018; Bauer et al., 2019; Hollowed et al., 2020). Where models differ in results will highlight where uncertainties in future scenarios may lie. If the results deviate greatly then those uncertainties may be considered to have a high priority to be resolved.

A method is available for overcoming model uncertainty to assist policy makers in determining suitable adaptation strategies for responding to future climate challenges while achieving ecosystem resilience. This method, often termed "management strategy evaluation" (MSE), is to evaluate the performance of those strategies under various climate scenarios within the social-ecological-climate models of the model ensemble that are considered to suitable reflect possible futures (A'Mar et al., 2009; Hollowed et al., 2020). Strategies that perform well across all plausible models, uncertainty and climate scenarios will be more likely to be successful in future real-world applications. Choosing suitable strategies will be dependent on trade-offs amongst social and ecological performance measures (objectives) and the degree to which risk of failure to meet those objectives can be countenanced, given the uncertainties. The success of this approach has been demonstrated in many fisheries applications. As yet, it has had only limited application across polar regions (but see Hollowed et al., 2020; Kaplan et al., 2021). Nevertheless, results from validated high-resolution ecological and social system models for the Bering Sea and Barents Sea provided much greater insight into the expected trajectories of climate change in the region (Hansen et al., 2019b; Holsman et al., 2020).

Despite these advancements in understanding, the future outcome of some key policy-relevant processes remain uncertain. These knowledge gaps tempered SROCC's confidence levels on impacts of climate change on polar marine social and ecological systems, and on the effectiveness of responses to climate driven change. Since the completion of SROCC, the availability of improved regionally downscaled physical biogeochemical models (e.g., Kearney et al., 2020) and observed responses of marine life to hazards attributed to climate change (e.g., North Pacific marine heatwaves) has further impacted key conclusions in AR6.

Ecosystem Process Error and Structural Uncertainty

Climate change affects multiple physical and biogeochemical processes that impact species throughout different life stages and at different spatial scales with sometimes attenuating (or amplifying) effects. These complexities are discussed in AR5 and SROCC. There are well-known tradeoffs associated with adding ecological realism into ecosystem models associated with adding parameter uncertainty and opportunities for model misspecification. For this reason, exploring outcomes using also regional multi-model ensembles that address different levels of mechanistic complexity is needed to reveal how these play out with respect to future conditions for fish and shellfish (e.g., Hollowed et al., 2020).

EBMs provide approaches for exploring ecological responses; however, such models should build upon well-developed theory and understanding of the ecosystem in question, which again is linked to availability of data for developing and evaluating the model. An analysis of different sources of uncertainty in long-term projections of fishing and climate change by EBMs (for the eastern Bering Sea) revealed that for some species structural model uncertainty dominated (Reum et al., 2020). Moreover, synergies across multiple EBMs with different underlying structural assumptions helped reveal potential short-term benefits of increased flexibility in catch allocation scenarios (Reum et al., 2020; Whitehouse et al., 2021). This also enhanced the ability of EBMs to provide guidance toward how to stabilize catches and forestall climate-driven declines (Holsman et al., 2020; Reum et al., 2020; Whitehouse et al., 2021).

Structural Error in Climate Models and Scenario Uncertainty

Structural uncertainties emerge in polar regions from global scale ensemble model projections as a consequence of results from different models in an ensemble being spread over a wide spatial range. Recent global scale ensemble model projections indicate substantial increases in total animal biomass toward 2100 under RCP2.6 (48%, intermodel SD 93.75%) and RCP8.5 (82.0%, intermodel SD 201.07%) in the Arctic (Bryndum-Buchholz et al., 2019; Lotze et al., 2019; Nakamura and Oka, 2019). This increase is partly due to increase in primary production fuelling the food chains, and partly due to increased biological rates with increasing temperatures. However, this does not take into account significant regional declines projected for the largest Arctic fisheries in the Bering sea, nor recent agreements to delay fisheries in the high Arctic until foundational information for sustainable management can be collated (i.e., 16+ yr moratorium on commercial fishing; Vylegzhanin et al., 2020; U.S. Department of State, 2021). Further, increased variability in biomass is also projected for polar regions, and of the models evaluated the polar regions had some of the lowest agreement across models in projected changes in biomass (Lotze et al., 2019). For the Southern Ocean there are no trends, but greater variability in both primary production and total animal biomass are projected under RCP2.6, while a 15% increase (intermodel SD 36.61%) in animal biomass is projected under RCP8.5 (Bryndum-Buchholz et al., 2019; Lotze et al., 2019; Nakamura and Oka, 2019). Thus, high inter-model variability in projections, combined with regional models projecting significant distribution shifts and declining productivity in key ecological and commercial species, demonstrate that the future development of polar marine systems and associated commercial fisheries are associated with high uncertainties (Griffiths et al., 2017; Klein et al., 2018; Hansen et al., 2019a; Tai et al., 2019). Despite these uncertainties, more evidence has emerged since SROCC's finalization demonstrating that sustainable fisheries practices within an ecosystem approach to fisheries management can stabilize fisheries and forestall negative impacts of climate change on some fish populations in the near term (Gaines et al., 2018; Free et al., 2020; Holsman et al., 2020; Reum et al., 2020).

Comparison of outputs from coarse spatial (and often coarse temporal) resolution global models with regionally downscaled simulations revealed systematic differences in the projections of future climate change impacts on high latitude systems in some high latitude regions. For example, scenarios from global models (often with annual time-steps) projected increased primary production and increasing biomass across functional groups in the Barents and Bering Seas, whereas downscaled models revealed seasonal differences in the timing of primary production and declining biomass trajectories for some functional groups (Hansen et al., 2019a; Holsman et al., 2020; Reum et al., 2020; Whitehouse et al., 2021). Emerging efforts to embed high resolution ocean model capabilities along the coastal shelf within global models with two-way coupling holds great promise for future IPCC assessments (Buil et al., 2021). Sustained support for high resolution modeling platforms and model intercomparisons of integrated oceanographic-ecological-social-economic models is needed to resolve the dynamic responses of coupled-social ecological systems to climate change (Holsman et al., 2019; Hollowed et al., 2020).

Ecosystem Resilience

A fundamental question on future polar marine ecosystems relates to their resilience; the capacity of the social ecological system to maintain the current state or return to some historical state following climate-driven change, whether the benchmark be pre-industrial or some time since. Could polar ecosystems cross a tipping point or threshold making it difficult to return and after which point productivity and stability are highly uncertain? A key source of uncertainty regarding the capacity of an ecosystem to recover is how and when future physical and biogeochemical changes in the ocean will trigger tipping points in ecosystem structure and organization that will limit recovery of the system to its former state (Frölicher et al., 2016; Frölicher and Laufkötter, 2018). Having capacity to project how the system will change, and whether tipping points may arise, is central to discussing prospects for polar ecosystems. SROCC moved significantly forward on this; a chapter was dedicated to "Extremes, Abrupt Changes, and Managing Risks" (Collins et al., 2019). However, that chapter to little degree dealt with polar regions and the knowledge summarized was not integrated into the polar chapter. Since SROCC the emphasis on abrupt changes and tipping points in the ocean has been pronounced, including work by Malhi et al. (2020), Turner et al. (2020), Degroot et al. (2021), and Heinze et al. (2021).

Spatial Heterogeneity

In SROCC, the authors intended to further examine the potential for spatial heterogeneity in Arctic and Antarctic ecosystem responses to projected climate change (first raised in AR5; Larsen et al., 2014). Unfortunately, such an inter-regional comparison was restricted by the lack of sufficiently resolved large-scale model output fields. CMIP5 output was found to be too coarse to differentiate between regions by capturing oceanographic conditions that have profound impacts on species distributions and interactions and food-web dynamics (e.g., sea ice distribution and edge blooms, seasonal stratification, Bering sea cold pool formation and extent; Kearney et al., 2020; Drenkard et al., 2021). Regional climate scenarios derived from down-scaled global climate scenarios and used to drive environmentally linked fish population models were included by Meredith et al. (2019), but were then only available for the Eastern Bering Sea (Hermann et al., 2019). Efforts since SROCC to downscale CMIP6 model outputs for coupled high resolution downscaled projects further underscored the importance, for polar regions in particular, to characterize impact variability and detail as well as adaptation efficacy at a regional scale (Hansen et al., 2019a; Drenkard et al., 2021).

Social-Ecological Responses

Dynamic couplings between regional social and ecological responses to climate change also deserve increased attention, as they can both amplify and attenuate climate impacts in complex ways. Global scale evaluations and models often are unable to capture important feedbacks and connections, leading to misspecification of impacts and adaptation response. For instance, realized harvest rates reflect complexities of management, economics, and regulatory structures that by design aim for ecosystem-based-management targets rather than maximizing the yield of individual stocks (MSY); total yield from a system is often lower than the additive sums of individual MSYs (or MSY proxies; Holsman et al., 2016, 2020; Link, 2018; Reum et al., 2020). Other ecosystem-based management targets include maintaining ecosystem structure, function and productivity (e.g., Norwegian Ministry of Climate and Environment, 2020), which may be less realistic under the ongoing directional climate change driven alterations of marine ecosystems. There is thus a need to adapt current management targets to the ongoing changes, and identify and evaluate relevant management strategies to reach these targets.

Disciplinary Imbalance in Scientific Priorities

Research regarding climate impacts in polar regions – particularly that for the Southern Ocean – is dominated by physical oceanography and cryospheric science with highly fragmented approaches to ecology. In terms of projections of climate impacts, the physical system out to 2100 is uncertain but well-circumscribed. By comparison, science on the effects of climate change on ecosystems and ecosystem services lags far behind (**Figure 1**). There is an urgent need for sustained support for long-term ecosystem data collections for Antarctic systems, linked ecological and socio-economic modeling at the circumpolar scale, and synthetic evaluations of cascading impacts of climate change, risks, and adaptation feasibility and effectiveness across Antarctic ecosystems and dependent industries. There have been few positive developments in Antarctic ecosystem research funding since AR5: in the Antarctic most long-term studies recently either lost funding or were greatly reduced. To address this major source of uncertainty in climate change assessments, research funding needs to be at the scale and scope to support understanding that can fully address policy and decision-making needs, i.e., cross-disciplinary and coupling ecosystem monitoring and high-resolution oceanographic and ecosystem models with social-ecological modeling. Such integrated research approaches take multiple years to establish but once developed can easily incorporate new projections and information. Sustained support (i.e., often through coordinated government investment in climate programs) to ecosystem based adaptation and mitigation is necessary to both develop and continuously update integrated approaches but the collective and holistic information that results is invaluable for identifying key climate risks, rates and magnitude of change and the effectiveness (and limits) of responses.

The challenges with establishment and proper coordination of large-scale ecological programs can, as noted above, partly be attributed to the costs associated with the spatial and temporal coverage required for ecological investigations beyond the local scale. There has been a scarcity of collaborative research initiatives at all levels to tackle the need for ecologically based studies of the Southern Ocean. During the last International Polar Year several scientific collaborations were initiated, including Census of Antarctic Marine Life (CAML) and Integrating Climate and Ecosystem Dynamics in the Southern Ocean (ICED). These have demonstrated the importance of networking and coordination in ecological research initiatives; such as Integrated Ecosystem Assessments (Levin et al., 2014) that are in development for Arctic ecosystems (Bering, Chukchi, Greenland and Barents Seas) as well as the Southern Ocean (MEASO, 2020). These IEAs are increasingly addressing the state and development of the polar seas as socio-ecological systems, assessing impacts and risks of climate change for species (Holsman et al., 2017), habitats and natural communities (MEASO, 2020), as well as on ecosystem services and coastal livelihoods (Holsman et al., 2020; Cavanagh et al., 2021). Such expansions are required to support the development of ecosystem-based adaptation and mitigation options to support management of polar systems under climate change impacts.

Policy Relevant Climate Assessments

Polar regions are a special case in IPCC assessments. They play a central role in the physical Earth System and have the potential for cascading positive feedbacks to Earth's climate system. Managing greenhouse gas emissions is the primary way that these changes in the Earth System can be moderated (IPCC, 2018). Changes in polar regions impact people the world over, not just from the perspective of physics and tangible services from within the region (Meredith et al., 2019; Cavanagh et al., 2021) but also from more distant biological and human connections with the regions (Murphy et al., 2021; Roberts et al., 2021). Adaptation may attenuate impacts of climate change to socio-ecological systems in the near term (see Simpson et al., 2021) but cannot protect the fundamental nature of cold- and ice-dependent marine ecosystems, which are projected to experience rapid and irreversible loss over the next century under high (and possibly moderate) emission scenarios. There is increasing and widespread agreement and evidence for this emergent finding across multiple lines of evidence, despite the uncertainty around the timing and rate of change. Yet, persistent uncertainties due to lack of temporal and spatial coverage, and gaps in cited resources, result in low or medium assessment confidence, potentially dampening the resonance of this critical finding. This causes a major challenge for delivering actionable science-based advice and for policy makers that must act now to address long-term climate risks.

Attribution of climate change impacts requires comparisons of impact event likelihood, through comparison of modeled or measured baselines, and/or projected to historical frequency of occurrence. For physical systems these baselines exist, enabling climate attribution, but for ecological and social systems such datasets are limited in space or time. Social-ecological systems integrate climate impacts across trophic levels and species and effects are lagged and often modulated through trophic interactions. This can make attribution of climate impacts difficult to detect when impacts are gradual or incremental, but has become more apparent with large-scale climate shocks (e.g., Huntington et al., 2020). Mismatched scales of ecological time series and seasonal and spatial gaps in information for biomass or rates of production for key ecosystem guilds (e.g., benthic detritivores) further challenge attribution. Coupling sampling and monitoring with multiple ecosystem models helps evaluate attribution and sensitivity of systems to climate impacts and is a near-term approach to improving ecological and social climate attribution.

ACTIONABLE RECOMMENDATIONS

We here present our most important recommendations with short explanations.

Development of a Scientific Framework That Moves From Assessing Future States With Attendant Uncertainties to Assessing Risks to Social and Ecological Needs Identified by Policy Makers

Assessment of climate change related risks requires development of frameworks and models that link across social and ecological dimensions, to allow evaluation of prognoses and management strategies under different scenarios. Management targets and solution options for socio-ecological systems in directional change needs to be developed, and they must capture key outcomes covering diverse societal needs and perspectives across stakeholder groups. Hence, a more inclusive approach to develop management targets and solution options is needed, both regionally, nationally and internationally. We therefore recommend both regional councils, polar nations, and international polar organizations (e.g., the Arctic Council, and CCAMLR) to establish such inclusive processes, building on diverse knowledge bases and perspectives from natural to social sciences, ITK, management bodies and other stakeholder groups. Also, in IPCC assessments, inclusiveness of e.g., ITK could be reflected in authorship and equity in contribution.

Greater Investment to Directly Assess Risks and Impacts of Climate Change on Polar Marine Ecosystems

While this is a well-worn statement in scientific literature, there is no doubt that much of the research to date used by the IPCC on the effects of climate change on polar marine ecosystems above biogeochemistry is opportunistic from outputs of other science programs. The direct, compound and cascading effects on society and ecology needs to be a directed effort.

Orientation of the Scientific Community to Progressing Repositories of Studies, From All Languages, on Climate Change Impacts in Polar Regions

Due to the strong information bias toward some regions (e.g., the Bering and Barents Seas, and the Western Antarctic Peninsula), any additional information on climate change impacts, particularly from other polar regions, will support the reduction of uncertainties and help support robust decision making for ecosystem management in a changing climate. Although charged with examining also such literature, IPCC authors may more easily overlook this. To support increased access and use of non-English literature, we recommend that IPCC WG chapter teams assessing polar regions should cover the central languages of the regions. If not possible to achieve for every assessment, the goal could be to balance representation over time. IPCC could also motivate the WG chapter teams to identify contributing authors that master relevant languages not covered by the chapter team. Finally, IPCC could identify and inform on search engines and best practices to find relevant literature from different regions and in different languages, and provide technical support to literature searches when needed.

Build Upon and Strengthen Ongoing Synergies Between Physical, Chemical, Biological, and Social Sciences in Assessing Social-Ecological Impacts of Climate Change, Their Root Causes, and Prognoses for the Future

For the full Assessment Reports, the IPCC could shift toward better integration between the three working groups as was done in SROCC (there WG1 and WG2). In recognition of the emerging integration of physical biogeochemical, ecological, social and economic research, the IPCC is encouraged to build formal conduits for information flow between working groups.

The authors have experience as lead authors in main AR reports and SROCC, one in both. For ecologists the latter approach was much preferred. Having physical and social scientists working alongside biologists on the same chapter, as in SROCC, was far more efficient and much more challenging with respect to moving our ecological knowledge to seriously take on the societal implications of climate change. One possibility is to have lead authors from one working group take part in the lead author meetings of the author two, serving as "ambassadors of knowledge." This will of course involve extensive traveling for these authors. Alternatively, one could have "regional ambassadors" who would be points of contact for key questions and discussions relevant to their working group.

CONCLUDING REMARKS

Our experience has shown that assessments on the future of polar ecosystems under climate change have been improving in recent cycles but there needs to be a phase shift in orienting the assessments to whole of system dynamics and impacts, including impacts on and risks to both social and ecological outcomes. We observe that there are a number of international initiatives that already provide the means for developing suitable polar ecosystem observing activities, providing there is support and impetus from national programs. A major gap, though with some valuable experience already on the table, is the development of a risk assessment framework for polar marine ecosystems that can utilize social and ecological whole of system models coupled to Earth System models. The development of such a framework can be common to both polar systems as they have similar requirements for their development and implementation. Lastly, we have identified a specific need for engagement across the broader scientific community and with national and international policy makers to develop repositories for the diversity of scientific information that will facilitate equitable assessments of the prognoses for polar marine ecosystems. Without an ongoing broad-based repository, polar system science will remain based only on those results that are readily disseminated and easily interpreted.

AUTHOR CONTRIBUTIONS

All authors listed have made a substantial, direct, and intellectual contribution to the work and approved it for publication.

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Identifying Barriers to Estimating Carbon Release From Interacting Feedbacks in a Warming Arctic

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The northern permafrost region holds almost half of the world's soil carbon in just 15% of global terrestrial surface area. Between 2007 and 2016, permafrost warmed by an average of 0.29°C, with observations indicating that frozen ground in the more southerly, discontinuous permafrost zone is already thawing. Despite this, our understanding of potential carbon release from this region remains not only uncertain, but incomplete. SROCC highlights that global-scale models represent carbon loss from permafrost only through gradual, top-down thaw. This excludes "pulse" disturbances – namely abrupt thaw, in which frozen ground with high ice content thaws, resulting in subsidence and comparatively rapid ongoing thaw, and fire – both of which are critically important to projecting future permafrost carbon feedbacks. Substantial uncertainty remains around the response of these disturbances to ongoing warming, although both are projected to affect an increasing area of the northern permafrost region. This is of particular concern as recent evidence indicates that pulse disturbances may, in some cases, respond nonlinearly to warming. Even less well understood are the interactions between processes driving loss of permafrost carbon. Fire not only drives direct carbon loss, but can accelerate gradual and abrupt permafrost thaw. However, this important interplay is rarely addressed in the scientific literature. Here, we identify barriers to estimating the magnitude of future emissions from pulse disturbances across the northern permafrost region, including those resulting from interactions between disturbances. We draw on recent advances to prioritize said barriers and suggest avenues for the polar research community to address these.

Keywords: Arctic, Boreal, disturbance, fire, permafrost, thaw, carbon, climate

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INTRODUCTION

Permafrost soils across the boreal and tundra biomes constitute the world's largest vulnerable carbon pool. Disturbance to this pool through ground thaw, and the subsequent magnitude, timescale, and form of carbon release into the atmosphere, will have a decisive impact on future climate change (Schuur et al., 2015).

Climate change drives disturbance of northern permafrost through a number of distinct processes (Kokelj et al., 2017) (**Figure 1**). These include gradual warming of the ground and subsequent thickening of the active layer; the upper layer of permafrost soils which undergoes seasonal thaw. This "gradual thaw" is projected to result in

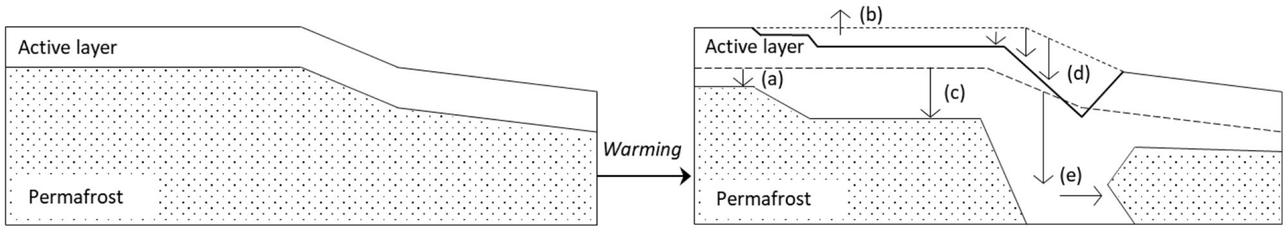

FIGURE 1 | Conceptual figure illustrating possible changes in permafrost environments due to ongoing climate change, including (a) active layer thickening; (b) combustion of soil organic matter during more frequent and severe fires; (c) accelerated active layer thickening following fire; (d) surface subsidence and development of abrupt thaw features, including in response to increased fire activity; (e) progressive, fire-induced thaw of the full permafrost table in more southerly regions of discontinuous permafrost, including through the formation and lateral expansion of taliks.

the loss of 24 to 69% of near-surface (upper 3m) permafrost this century (likely range; IPCC, 2019). Estimates of carbon release resulting from gradual thaw on the same timescale range from 30Pg C to 150Pg C (Natali et al., 2021); equivalent to the cumulative emissions through 2100 from Japan to those of the United States through 2100, at their current rate of emissions (UCS, 2020).

Gradual thaw can be likened to a "press" disturbance; a disturbance with a comparatively low magnitude and long duration (Jentsch and White, 2019). Currently, only gradual permafrost thaw is represented in Earth system models (ESMs). This excludes important, relatively slow press disturbances such as the formation of subsurface layers of perennially thawed soil known as taliks, which can drive additional thaw (Devoie et al., 2019), and some gradual thermokarst processes (such as lowland thermokarst lakes), which result from uneven ground subsidence following thaw.

However, climate change is driving not just an acceleration of press disturbance, but also an increase in some types of "pulse" - abrupt, high magnitude per unit duration - disturbance across permafrost regions. This is of concern as pulse disturbances can in many cases result in substantial carbon emissions.

Notable among these pulse disturbances is abrupt permafrost thaw. Abrupt thaw is a collective term encompassing a range of thermokarst processes, where degradation of hillslope permafrost with a high ice content results in comparatively rapid, high magnitude ground disturance and the formation of features (e.g. retrogressive thaw slumps, Kokelj and Jorgenson, 2013). The term also includes the development of some lowland thermokarst landforms, such as thermokarst wetlands and lakes; although some thermokarst features form gradually over decades, and therefore may be better described as press disturbances. While gradual thaw proceeds top-down over years and decades, abrupt thaw can expose several meters of permafrost on a timescale of days to years.

Of the pulse disturbances considered here, fire affects the largest area of the Arctic-boreal zone (Stocks et al., 2002; van der Werf et al., 2017). The incidence of fire has already increased in boreal and tundra biomes since the mid-20th century (Kasischke et al., 2010; Hanes et al., 2019), with evidence suggesting that intensifying fire regimes are linked to exceptionally high fire emissions in recent years (Veraverbeke et al., 2017; Walker et al., 2018; McCarty et al., 2020; Scholten et al., 2021). While some ESMs include fire, they exclude combustion of soil organic matter which is the dominant means of carbon loss at high latitudes. Further, fire can accelerate permafrost thaw following combustion of insulating soil organic layer, both by initiating or accelerating gradual thaw and by promoting abrupt thaw processes in ∼20% of the high-latitude region which is prone to abrupt thaw (Olefeldt et al., 2016; Gibson et al., 2018).

Although pulse disturbances occur comparatively infrequently in time and space, they can have impacts equal to those of press disturbances. For example, across the Arctic, abrupt thaw could result in radiative forcing equal to that of gradual thaw, despite affecting less than 20% of the permafrost region (Turetsky et al., 2020).

The body of literature addressing carbon release from pulse disturbances is growing, and in some cases, notably for fire, is already substantial. However, the stochastic nature of pulse disturbances, the complexity underlying both their impacts and responses to climate change, and the challenges inherent in acquiring data across high-latitude regions, are still major obstacles to incorporating these important processes into comprehensive assessments of future carbon release from the northern permafrost region. Here we reflect on these obstacles, with the objective of highlighting priority issues for future research and approaches which - we argue - represent the most efficient means of addressing these.

BARRIERS TO ESTIMATING CARBON LOSS

Of the disturbances considered here, recent carbon loss from fire can be assessed with the most confidence. A number of gridded datasets derived from satellite data report burned area and fire emissions at annual or greater temporal resolution at regional to pan-Arctic scales (Randerson et al., 2015; Veraverbeke et al., 2015; Giglio et al., 2018; Otón et al., 2019; Dieleman et al., 2020). Links with climatic change are also well established, with warming, drying, and increased lightning strike rate all implicated in increasing fire activity (Walsh et al., 2020; York et al., 2020; Chen et al., 2021; McCarty et al., 2021).

However, projected increases in burned area, a primary determinant of fire emissions (Veraverbeke et al., 2015), range from less than 50% to more than 150% per °C of global warming, with similarly wide ranges reported even for well-studied regions such as Alaska (7–93% $^{\circ}C^{-1}$) (estimated from: Euskirchen et al., 2009; Eliseev et al., 2014; Genet et al., 2018). The breadth of these ranges demonstrate that future fire activity remains highly uncertain, due both to the complexity of the underlying processes and the challenge of scaling these across large regions (Kitzberger et al., 2017; Boulanger et al., 2018).

In contrast to fire, there is no pan-Arctic assessment of abrupt thaw incidence. This is partly due to the difficulty of detecting abrupt thaw features - which can be comparatively small and, in contrast to the thermal signatures associated with fire, lack a signal that is easily detectable through moderate-resolution satellite remote sensing. Recent studies illustrate that even High Arctic permafrost can rapidly undergo abrupt thaw in response to warming (Farquharson et al., 2019; Jones et al., 2019; Lewkowicz and Way, 2019). However, while local-scale observational studies provide critical insights into (for example) the drivers of different thaw trajectories, an ongoing reliance on local-scale studies alone to quantify abrupt thaw incidence across large and heterogeneous Arctic regions means that the links between climatic change and abrupt thaw rates remain poorly constrained.

Similarly, although regional-scale modeling demonstrates that abrupt thaw can have decisive consequences for carbon vulnerability (Nitzbon et al., 2020), comprehensive assessments of its impact on carbon fluxes are scarce, particularly regarding lateral fluxes, which can account for more than half of total carbon loss (Plaza et al., 2019). The one existing pan-Arctic estimate suggests that abrupt thaw-driven net carbon losses could equal 40% of those for gradual thaw (Turetsky et al., 2020). However, data scarcity remains a limiting factor in this first-order approach.

Quantification of post-fire thaw also relies heavily on single-site studies, often with less than a decade of observational data reported (but see e.g. Jafarov et al., 2013; Gibson et al., 2018). For tundra ecosystems, where the largest proportional increases in fire regimes are expected (Chen et al., 2021), no long-term monitoring of post-fire active layer thickness has been reported to date (Holloway et al., 2020). Therefore, our understanding of post-fire thaw trajectories, and especially their spatial and geographical variation, remains limited.

The obstacles posed by a lack of observational data are compounded by limitations in current modeling approaches. The spatial resolution at which global scale models operate is much coarser than stochastic, fine-scale and highly heterogeneous pulse disturbance processes. Moreover, simplified model structures do not lend themselves to mechanistically simulating these disturbances. For example, ESMs typically employ one soil column per grid cell, and a "big leaf" representation of vegetation; effectively collapsing complex eco-physiological, biogeochemical, and geomorphological processes across thousands of km^2 into a single site.

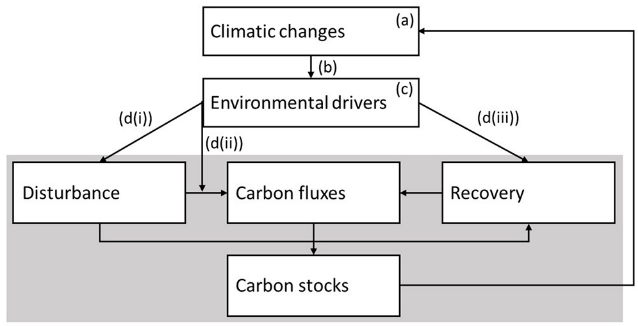

FIGURE 2 | Simplified framework for predicting carbon loss from pulse disturbances under climate change. Predictive understanding relies on a foundational understanding of how the incidence of and recovery from disturbance affects carbon fluxes, and how those carbon flux changes translate to longer-term changes in carbon storage (processes shaded in light gray). Predicting future carbon loss from disturbance under climate change scenarios then requires (a) predictions of climatic changes; (b) mechanistic understanding of how these variables drive (c) key environmental variables that influence incidence of the disturbance in question, and data describing variation in these variables across tundra and boreal biomes; and (d) a mechanistic understanding of how these variables affect (i) disturbance characteristics such as frequency, duration, and spatial distribution; (ii) the severity or intensity of impacts on ecosystem carbon fluxes and (iii) characteristics of recovery such as duration and final state.

PRIORITIZING FOR BETTER PREDICTIONS

Predicting future consequences of pulse disturbances is complex. It requires predictions of climatic changes, identification of the key environmental drivers (including variables describing environmental context, such as those related to soil characteristics, topography and drainage) which govern the incidence of, and recovery from, disturbance, and mechanistic understanding of how climatic changes affect those drivers. Data describing geographical and spatial variation in key environmental drivers, and mechanistic understanding of how those drivers affect the incidence, impacts of, and recovery from disturbance are also required to apply predictions across large spatial scales.

We propose a conceptual framework (**Figure 2**) to describe these data and knowledge needs. Below, we use this framework to identify priorities to deliver near-term improvements in predicting carbon loss from pulse disturbances. In doing so, we emphasize the need for improvements in: (a) mechanistic understanding of how environmental variables drive disturbance impacts (**Figure 2**d); and (b) gridded data describing these driving variables (**Figure 2**c). "Mechanistic understanding" here refers to a predictive understanding of the means through which variation in a given environmental variable drives variation in the impacts of disturbance. "Driving variables" here refers to variables with causal influence over disturbance impacts. Together, improvements in (a) and (b) can deliver better understanding of how heterogeneity across the Arctic currently influences the incidence and consequences of pulse disturbances, and therefore how environmental change will influence pulse disturbances in the future. Addressing each of these concepts (a & b) respectively, we provide specific recommendations and priorities for future research.

Identify Mechanisms

Identifying and quantifying causal processes can highlight variables that influence disturbance characteristics in predictable ways. These variables can then guide ground or remote data collection and/or inform predictive modeling approaches. For example, recent work has highlighted that "bottom-up" drivers, such as fuel availability, can exert a stronger influence over fire emissions compared to "top-down" drivers (such as fire weather; Walker et al., 2020). In this case, site-level drainage played an outsized role in determining fuel availability and carbon emissions, yet large-scale estimates of drainage do not exist. It follows that better quantifying drainage conditions should be a high research priority for understanding and predicting the impact of intensifying fire regimes on carbon cycling.

A mechanistic focus has strong potential to reduce uncertainty around carbon loss due to post-fire permafrost thaw. It has been shown that post-fire organic layer depth is a decisive factor in determining subsequent thaw trajectory (Jafarov et al., 2013). And since belowground carbon pool size, vegetation type, and soil moisture influence organic layer combustion, these variables are likely significant drivers of post-fire thaw (Minsley et al., 2016; Walker et al., 2020).

Mechanisms driving ground thermodynamics are relatively well-understood (Jafarov et al., 2013), allowing robust site-level estimates of gradual ground thaw (e.g., Garnello et al., 2021); yet, pan-Arctic assessments are limited by lack of robust gridded datasets (e.g., snow depth and density). While ground ice content is the primary determinant of ground subsidence following thaw, abrupt thaw processes are also strongly influenced by landscape characteristics such as surface moisture content, hydrological connectivity, soil type, topography, slope and deposit stratigraphy (Shiklomanov and Nelson, 2013; Lara et al., 2015; Olefeldt et al., 2016). Further, the initiation and progression of abrupt thaw may be exacerbated by other pulse disturbances, including fire (Raynolds et al., 2013; Baltzer et al., 2014; Lewkowicz and Way, 2019; Christensen et al., 2021). A causal, quantified understanding of how these landcover characteristics – and interacting disturbance types - influence abrupt thaw processes could therefore inform improved pan-Arctic carbon predictions.

Another priority area for improved mechanistic understanding is post-disturbance changes in vegetation characteristics, such as increased productivity (Reichstein et al., 2013), or dominance of deciduous species (Mack et al., 2021). It is often suggested that these changes may to some extent "offset" soil carbon loss; but long-term, integrated understanding of disturbance-driven changes in carbon uptake is needed to assess how realistic this is – particularly as individual disturbance events can have large, near-term impacts on net carbon loss (Abbott et al., 2016). For example, emissions from the Anaktuvuk River fire, which burned 1,039 km^2 of Alaskan tundra in 2007, were equivalent to annual net carbon uptake by the entire tundra biome (Mack et al., 2011). Further, fire has knock-on impacts on permafrost by changing ground surface albedo and removing the vegetation and organic soil that insulate ground temperatures and permafrost from warm summer air temperatures; e.g. via shading and the insulative/conductive capacity of dry/wet moss (O'Donnell et al., 2009). These effects can promote gradual, top-down permafrost thaw over a number of decades, as well initiating the development of thermokarst features, including larger-scale abrupt thaw features (e.g. retrogressive thaw slumps, Liu et al., 2014; Jones et al., 2015). However, the trajectory of post-fire permafrost thaw depends on processes such as the depth of combustion of the soil organic layer and the trajectory of post-fire vegetation recovery; which are in turn mediated by variables such as topography, soil moisture status and ground ice content (Jafarov et al., 2013; Nossov et al., 2013; Gibson et al., 2018).

A process-based approach is also an avenue for addressing nonlinearity. Recent work suggests high-latitude fire regimes may be undergoing a step-change in response to climate change, featuring an abrupt advance in the timing of the fire season. This advance is thought to indicate an increasing role for holdover or "zombie" fires which persist throughout the winter through sub-surface smoldering, before re-commencing their spread the following spring (Scholten et al., 2021). Similarly, the last few years have seen increased burning of traditionally fire-resistant landscapes such as wetlands (McCarty et al., 2020). These phenomena highlight the value of process-level understanding in addition to observations of net impacts; in a rapidly changing and no-analog climate, responses that are unfamiliar and perhaps unexpected are not unlikely. While some changes will inevitably remain unpredictable, a quantified understanding of the variables and thresholds driving key processes such as fire ignition, spread and smoldering behavior may highlight the potential for non-linearity.

Recommendations

While acknowledging that many knowledge gaps exist, here we have identified a sub-set of the processes that determine the net impact of abrupt thaw, fire and fire-mediated thaw on carbon balance that are missing from models, and for which it is feasible to develop mechanistic understanding.

First among these is surface moisture dynamics. Although surface water can significantly alter permafrost temperatures (Langer et al., 2016), how moisture status influences processes such as fire ignition remains poorly quantified, as does its influence over pathways of subsequent carbon release (Plaza et al., 2019). The importance of these processes will only increase as Arctic weather becomes increasingly rain-dominated and permafrost thaw results in complex hydrological change (AMAP, 2017).

A second priority is process-level understanding of vegetation responses and carbon re-accumulation following disturbance (Pizano et al., 2014). For abrupt thaw in particular, these responses are rarely monitored and poorly understood, despite the likelihood of substantial consequences for the net carbon impacts of disturbance (Hugelius et al., 2020). Long-term monitoring of post-disturbance vegetation processes remains a critical avenue for future research.

In addition, a systematic, mechanistic approach to future work on abrupt thaw processes should aim to provide causal insight into how site-specific variables – such as soil and hydrological characteristics, in addition to ground ice – determine the progression and net carbon flux consequences of abrupt thaw (e.g. Kokelj et al., 2017). Such insight, when paired with mapping/quantification of those key variables (see next section) could effectively address this core obstacle to comprehensive estimates of future high-latitude carbon loss (Turetsky et al., 2020).

Similar to ongoing developments in vegetation models (Fisher et al., 2015; Fisher and Koven, 2020), representing patch dynamics (or adaptive tiling) of abrupt permafrost thaw could help bridge the spatial gap between ESMs and abrupt permafrost thaw features. Although spatial resolution is improving, it is unlikely global models will be able to directly simulate these features at their native resolutions (as small as 5–10 m) in the foreseeable future. When developing new algorithms, modelers would also benefit from new benchmarks on abrupt thaw, particularly in terms of spatial distribution, rates of change, stabilization/recovery, and impacts on hydrology and carbon fluxes.

Quantify Drivers

Section Identify Mechanisms emphasizes the importance of identifying variables with causal influence on the net impacts of disturbance. Such variables may then be used in remote sensing and/or predictive modeling approaches to better constrain estimates of future disturbance impacts. For example, parameterizations for abrupt thaw (similar to those developed in Turetsky et al., 2020) could be based on grid cell functions of ground ice, topographic indices, soil properties, and other environmental and climate drivers. Mapping driving variables such as these over space is particularly valuable at high-latitudes, where geographical and spatial heterogeneity enhance the complexity of developing pan-Arctic predictions (Myers-Smith et al., 2020; Virkkala et al., 2021).

However, this approach requires that identified driving variables can be quantified at a sufficient resolution to resolve landscape heterogeneity. Data acquisition across remote Arctic regions remains challenging, and while remote sensing presents a solution for measuring some vegetation characteristics, satellite data can rarely be used to directly observe belowground variables. As a result, many variables fundamental to permafrost processes and belowground carbon cycling remain poorly quantified (Parker et al., 2021).

Notable among the variables which remain poorly-described and have substantial potential to reduce uncertainty is ground ice content: a primary determinant of susceptibility to thermokarst. Other similarly important variables include the depth of the soil organic layer, which plays an important role in determining permafrost thermal dynamics, and those relating to surface and soil moisture, which impact the susceptibility to disturbance, as well as the subsequent form and magnitude of carbon release (Plaza et al., 2019). Mapping these belowground variables could help facilitate the up-scaling of existing observational studies and constrain process-based predictive approaches.

Recommendations

Section Quantify Drivers highlights the underrepresentation of belowground variables in gridded data repositories as a limiting factor in understanding and projecting high-latitude disturbances. Description of belowground variables, and particularly those which are most critical to projecting permafrost changes, is therefore a primary need. Specifically, we recommend gridded data products describing ground ice content, organic layer thickness, and surface moisture at spatial resolutions that facilitate comparison with on-the-ground data as priority research needs.

On a practical level, the inherent difficulty of measuring belowground variables remotely and/or over large areas is a core reason for the comparative lack of data describing them. Therefore, alongside pursuing promising approaches to improving remote sensing of belowground variables, such as use of multi-frequency radar, alternative systematic approaches to collecting these data at a sufficient resolution are required. In particular, the success of initiatives such as ITEX (International Tundra Experiment; Henry and Molau, 1997) highlight the potential for centralized networks to guide efficient collection of comparable datasets at large geographic scales., Centralized protocols may extend these benefits further (Parker et al., 2021), particularly where these a transdisciplinary approach; integrating the knowledge and data needs of the ecosystem ecology, geomorphology and modeling communities.

DISCUSSION

Pulse disturbances present a considerable obstacle to accurately predicting future carbon dynamics, impeding projections of future climate feedbacks from the permafrost region. The body of literature addressing this issue is rapidly increasing, and includes identification of emergent relationships that may help constrain future disturbance rates and aid in up-scaling their impacts (e.g., Walker et al., 2020), as well as useful first-order, pan-Arctic estimates of those impacts (e.g., Turetsky et al., 2020). However, the availability of data required either for constraining such estimates, or scaling mechanistic insights across Arctic regions, remains insufficient. In the case of abrupt thaw and post-fire thaw, the mechanistic insight required to scale these data also limits our predictive capacity. Further, issues such as the disparity between fine-scale pulse disturbances and the relatively coarse spatial scale of existing global scale models, continue to present the modeling community with barriers beyond data availability to integrating these important processes.

We argue that future work should prioritize mechanistic understanding, focussed on identifying key driving variables, and improvements to the availability of data describing those variables. This represents an efficient approach to reducing the uncertainty associated with these important pulse disturbances.

However, the likely relevance of pulse disturbance impacts to near-term mitigation decisions demands not just an efficient approach to improving scientific understanding, but the maximal use of existing information alongside these efforts. This could encompass, for example, novel modeling approaches such as data assimilation (Scholze et al., 2017; Fox et al., 2018), and the incorporation of non-traditional data types such as expert assessments (Abbott et al., 2016; Sayedi et al., 2020). There is broad acceptance within the research community that pulse disturbances will significantly – albeit to an unquantified extent - impact net carbon loss from the permafrost region (Natali et al., 2021). It is therefore appropriate to prioritize effective dissemination of the risks associated with this, alongside longer-term efforts to facilitate a comprehensive, accurate estimate of the net carbon impact of pulse disturbances under ongoing climate change.

DATA AVAILABILITY STATEMENT

The original contributions presented in the study are included in the article/supplementary material, further inquiries can be directed to the corresponding author/s.

AUTHOR CONTRIBUTIONS

RT, SN, and BR designed research. RT, SN, BR, and EM performed research. RT wrote the paper. RT, SN, BR, EM, and TG provided discussion of themes, ideas, content, and editing. All authors contributed to the article and approved the submitted version.

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