Original Article

Studying the Impact of Physicochemical Parameters in Marine Sediments

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Abstract

Sedimentation refers to a systematic process of particle transport to, placement on, withdrawal from, and maintenance on the seabed. Sediment is a naturally occurring component of aquatic habitats. However, the quantity and features of sediment have an impact the marine ecosystems' physical, chemical, and biological integrity. Deep sediments are the nation's most significant geological deposits. They serve as a record of our planet's reaction to circumstances at the sea surface, from which the majority of the material comes. They are also crucial aspects of energy flow through food webs and contribute to the operating of marine ecosystems. In addition, they are essential in sedimentary procedures such as carbon cycling and, nutrient and, waste malfunction, and abolishment. By studying the living and non-living variables contributing to the coastal system, the study ofstudying these sediments can be a tremendous asset for studying climate change and anthropogenic consequences (pollution associated with chemicals or metals)), and it can be a biological indicator of climatic changes anthropogenic consequences. So, the purpose of this review is to focus on the study of these sediments that are attributed to the impacts of urbanization along the marine coast by evaluating the Physicochemical characteristics of marine sediments that affect the marine bio-communities negatively or positively, and thus assessing the climatic changes and anthropogenic consequences that are associated with human health, industrial, and economic fields.

Keywords: Physicochemical characteristics; marine sediments; bio-communities; ecosystems; minerals, Heavy metals

Introduction:

What are marine sediments?

Sediment refers to the small particles of rocks and other materials that sink to the ocean floor and eventually end up and accumulate on it. Sedimentation refers to a systematic process of particle transport to, placement on, withdrawal from, and maintenance on the seabed[3]. Sediment is a naturally occurring component of aquatic habitats. However, itsthe quantity and features of sediment have an impact on the aquatic ecosystems' physical, chemical, and biological integrity[4]. Sediment refers to the tiny particles of rocks and other materials that sink to the ocean floor and eventually end up and accumulate on it. Sediments can be found in all areas of the seafloor, which come from a variety of sources and accumulate in widely divergent amounts. Globally, ocean sediments range from 1 to more than 15 km thick; in places of more significant accumulation, near river mouths where sediment discharge is vital, as well as passive margins near continents where seafloor sediments have accumulated over millions of years [5].

The seafloor is made up of basalt formed at mid-ocean ridges and is covered in most locations by layers of sediment; any collection of loose particles, such as sand on a beach, mud at the bottom of a lake, or gravel on a riverbed. Marine sediments are near continents cover about 25% of the seafloor but may make up nearly 90% of all sediment deposits in value and volume[6]. Marine sediment combines any insoluble material deposited on the seafloor, transported from land areas to the ocean by the wind. Primarily rock, soil particles, ice, rivers, and originating from the stays of marine organisms, continental erosion, biological productivity, hydrothermal vents, submarine volcanism products, cosmic debris and materials from outer space, and chemical precipitates from seawater that collect on the seafloor[7].

Marine sediments cover much of the ocean's approximately 70% of the Earth's surface, making it the world's largest single ecosystem in terms of area. Provided their extensive coverage, it is not remarkable that sediments are species-rich systems, especially given that sediments range from cobble to fine clays. These sedimentary habitats range in salinity, temperature, and exposure from tidally exposed areas to enormous depths with no light, little variability in the environment (env-t), and pressure exceeding 100 times that at sea level[8].

Several previous studies regarding the study of living organisms, biological diversity, the impact of population activity, and pollution of water bodies. However, studies disregard the physicochemical features, the quantity of minerals or their absence, and the extent to which these properties have an influence.

Numerous water bodies worldwide lack appropriate research on their physicochemical qualities, the presence of precious metals in their bottoms, and even the extent and degree of contamination by heavy metals, as well as the impact of this on their physicochemical properties and biodiversity. Understanding and studying this enables the extraction of minerals and the knowledge of the impact of this nature on the aquatic environment along the Saudi Arabian Red Sea coast, which has a significant economic return, particularly in the countries' search for natural sources of alternative energy, mineral wealth, and fisheries.

Types and classification marine sediments:

Wind, water depth, and ocean circulation play a vital role in the transport, deposition, and maintenance of different sediment types on the seafloor[9]. It becomes evident that many processes contribute to the formation of marine sediments. They are carried by glaciers, rivers, the wind, and organism skeletons and shells. Rivers and glaciers drain huge quantities of sediment from continents, which are then blown out to sea by the wind. Organisms in the surface water produce a steady stream of skeletal materials, which fall to the seafloor. Chemical reactions from other sediments on the spot. The process of altering the forms of sediments, known as autogenesis, has a significant impact on the final constituents of marine sediments [10].

Sediment categorization is necessary to separate the examined sediments into multiple classes based on their dimensional ranges and to compare different types of sediments. Because there are several classification methods, the best one varies depending on the study's goal, the type of sediment being studied, and the accuracy of the analysis [2].

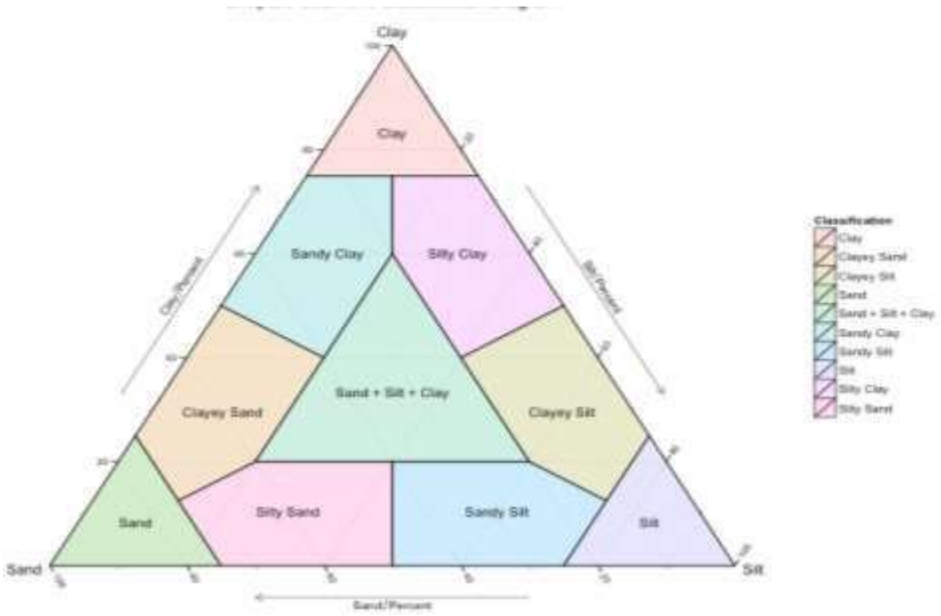
Some of the most common sediments' features to classify are their texture, composition, and origin[11]. Because coarser sediments are found near the coast, and at shallow water depths, while finer sediments rise with rising bathymetry, diverse aquatic env-ts are characterized by different sediment types[12]. There are some exceptional cases because fine sediments can be discovered not only in deep water env-ts andbut also in confined marine and transitional settings. In contrast, sandy sediments can be founddiscovered offshore due to low environmental energy due to glacial low-stand or gravitational processes during the Quaternary[13]. Meteorology, climate, hydrodynamics are The main factors affecting the distribution of sediment particles once supplied to the marine environment, where they undergo repetitive resuspension, deposition, and transport over the continental shelf and slope before getting definitively deposited, are meteorology, climate, hydrodynamics, and bottom morphology[14].

Sediment texture; a very stable characteristic related to total sand, silt, and clay proportions contents, influences soil biophysical properties[15], and lots of important biotic and abiotic aspects in marine environments, inconsequence, sediment texture is considered a critical factor in many fields. Diversity and composition of macro-and meiobenthic communities, organic carbon, and distribution of polluting substances are strongly limited by sediment grain size[16]. As a result, sediment texture involves determining the ecological features of benthic habitats and/or for environmental (env-tal) evaluations of marine coastal areas impacted by anthropogenic impacts, and expertise of sediment texture is a major concern in marine environmental research [2].

Sediment texture is examined through several variables, of which grain size is the most important; because it is the most basic physical property that represents the conditions in which the sediment was deposited. Some of the areas of research that benefit from sediment grain size information include sediment pollutant research, benthic society ecology, seismic studies, remote sensing questionnaires, and beach nourishment [2].

Sediment grain-size data provide a wealth of information on sediment transport and sedimentation, and they are crucial for deciphering trends in surface processes associated with the hydrodynamic characteristics of the sedimentary regions in question, as well as transportation and deposition circumstances[17]. Various grain sizes maintain the commitments of distal and proximal sediment sources, the factors of which are intimately linked to alluvial and coastal aggradation/degradation processes [18]. Other important applications of grain size data are for: a) Metal and mineral content concentrations are subject to standardization techniques [19], b) seismic unit relationship; recognized in profiles with corresponding stratigraphic, or for remote-sensing data calibration [20], c) Because the borrow area's features must be more or less similar to those of the native beach, coastal management, particularly beach nourishment, necessitates sediment description [21].

Many seabed substances have a distinct sediment composition. Sediment grains are a range of grain diameter sizes that have been cate as grades (fine and medium sand) and aggregates (mud, sand, and gravel) in naturally occurring varieties [22]. Particle size is used to classify sediments, ranging from the smallest clays (diameter less than 0.004 mm) to the biggest boulders (diameter greater than 256 mm). Circumstances with much energy, including strong currents or waves, cause only the larger particles to settle, while the finer particles are carried away. Under lower energy circumstances, smaller particles will settle out and shape finer sediments [23]. Shepard ternary categorization is one of the most frequently used techniques to classify marine sediments; either the original (based on the relative abundance of sand, silt, and clay fractions) or the more recent (depending on the relative proportions of these three constituents) with the identification of 10 major sediment textural groups [24], or the modified, which is beneficial when a huge proportion of gravel is documented, as the vertices comprise gravel, sand, and mud (silt + clay) for contamination evaluation, attention to the significance of clay minerals in the accumulation of heavy metals and organic compounds [2]. The percentagesThe %ages of gravel, sand, and mud are plotted on a triangular diagram for this categorization [24].



Sediment classification diagram by SEDCLASS program [2]

Another way to categorize sediments is via the ability to run water to move sediments or sorting, i.e. the degree of uniformity of grain size, concerning the size. More extended transport or exposure to waves and currents causes sediments to be sorted by shape and size. The quantity of sorting is determined by the energy circumstances and the length of time that stream currents or ocean waves interact with the particles[25]. Longer transport or exposure to waves and currents causes sediments to be more sorted by shape and size. Along with the transporting medium's power generation, particles are also sorted according to density.The sediment is well-sorted Ifif all of the particles are the equivalent in size and shape, as in beach sand, it is well sorted, whereas it is poorly sorted if they are all of the various sizes, it is poorly sorted. Glacialas in glacial deposits, for example [26]

Sediment texture is also classified based on the maturity of the particles or the length of time they have been carried by water or other matrices[27]. The degree to which particles are adjusted, the extent to which they are sorted, and the sediment structure reflect maturity. The rounder the particles become due to particle irritation over time, the more mature the sediment is. A greater level of sorting illustrates maturity because tiny particles cleaned away over time, and a given amount of energy will proceed particles of similar size over the same distance [27]. The greater the quartz content, particularly in sediments obtained from rock particles, the older and more mature the sediment. Quartz is a prevalent mineral found in terrestrial rocks known for being extremely hard and abrasion -resistant. Particles made of other materials wear away over time, leaving only quartz behind. Beach sand is mature sediment primarily composed of quartz particles that are rounded and well-sorted into roughly the same size [11].

Sediments are shown to vary in their sizes, shapes, chemistry, and density [28]. Marine sediments are also classified on the basis of the origin or source material and grain size[29], which are easily identified from their mineral composition. The occurrence of rock-forming minerals such as quartz and feldspar indicates their origin from land rocks. The presence of silica (SiO_2) and calcite ($CaCo_3$) is attributed to organism-derived compounds. Numerous minerals also precipitate from oceanic water. In this context, sediments are in four main categories; lithogenous/terrigenous, biogenous, hydrogenous, and cosmogenous[28]. Litho- means stone, and generous is derived from the suffix-gen meaning produced. Lithogenous means derived from stones or rocks [30]. Most marine sediments, especially near the shore, are of this type. Lithogenous sediments originate from erosion of land, volcanic eruptions, and blown dust which break apart rocks on ground[31]. Moreover, WeatheringAlso, the weathering of minerals on continental lithospheric rock bodies provides the input for lithogenous sediments in the oceans;, such as quartz sands, clays and muds, but clays are the major products of mineral weathering reaction on land. Sandstones, mudstones, and shales are the main types of rocks formed from terrigenous sediments that mainly exist as continental masses and as volcanic islands inside the oceans [11].

Because they are made up of rock fragments (originating from other rocks), lithogenous sediments are also known as clastic sediments. A clast is a Greek word for a rock fragment[31]. They are also known as terrigenous sediments because they originate on landmasses (terra meaning earth or land) and are generated from preexisting rock. These sediments are introduced by flowing water, wind, or occasionally ice[30]. Because larger sediments require more energy to move, they rarely travel long distances, but erosion keeps breaking them down into smaller particles. Because smaller residues need less energy to move, they can travel a greater distance. Most of these terrigenous sediments eventually end up in the ocean. Rivers and streams carry a significant part of sediments into the ocean, in which they settle out due to the reduced force of water. Larger rocks settle near the shore, but underwater landslides transport them much further out onto the seafloor[1]. Minor silt and clay particles are carried thousands of miles by ocean currents, eventually creating the deep ocean's abyssal clay or red clay layer. Because these sediments are more abundant near continents, particularly near river mouths and in very deep oceanic areas, their structure is influenced by the climate and weathering in the source region [30].

Biogenous sediments are made up of insoluble marine organisms stays. The remains of shells or fragments from shelled aquatic creatures and corals make up a large portion of these sediments in several areas where the water is shallow. These sediments are made from microscopic shells deposited by tiny plants, animals, and planktons that live on the water's surface and ultimately make their way down to the ocean floor in deep seas where there is not such a high concentration of these life forms[32]. Carbonate salts are produced by organisms corals, molluscs, and foraminifera. These are placed on ocean floors as calcite layers, resulting in limestone [7]. Other biogenic rocks include stromatolites, the flint nodules present in chalk, and coal. Also, bones, teeth, and insoluble organisms' ruminants are continuously deposited inside the oceans. The most common chemical compounds found as biogenous sediments are calcium carbonates; due to the presence of microscopic diatoms, animals; radiolarians, organisms; like foraminifera, are responsible, and silica. Growing Algae and coral reefs are the outcomes of the biogenic sediment formation[33].

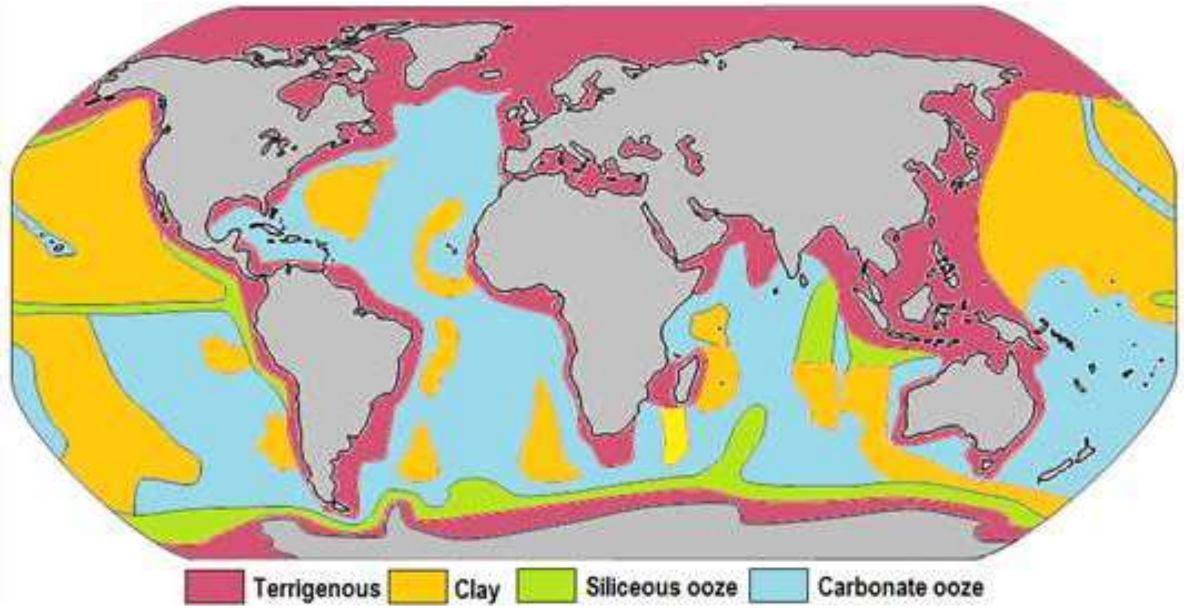
Three essentialimportant processes monitor the distribution of biogenous sediments: 1) their rates of production in surface waters; this is linked to the growth of marine algae (the base of the oceanic food chain), which is monitored primarily by nitrogen and phosphorus availability; 2) their rates of dissolution in deep-ocean water; where this area is undersaturated with calcium carbonate and opaline silica; and 3) their rates of dissolution in deep-ocean water; . As a result, biogenic particles tend to dissolve as they settle through the water column and settle on the seafloor, and 3) other sediments dilute them on the seafloor. Calcareous, as calcareous and siliceous elements are diluted to less than 30% of overall sediment in high terrigenous sediment input areas[11].

Coastal areas have very high primary manufacturing capacity due to the abundance of biogenous deposits. To be named an ooze, the sediment must contain more than 30% biogenous material (any sediment containing more than 30% microscopic skeletal debris is termed an ooze; this refers to the origin of the sediment, not its consistency) [31]. Even in constructive coastal areas, lithogenous input is so high that biogenous materials are overwhelmed, and the 30% threshold is not met. So lithogenous sediments influence coastal areas, whereas biogenous sediments are more prevalent in pelagic environments with little lithogenous input [1].

On the contrary, hydrogenous sediments originate from chemical precipitation reactions within the supersaturated seawater, when the dissolved materials (3.5 % by weight) precipitate out and form solid particles on the seafloor [32]. Typically, these reactions are stimulated by an alteration in the circumstances, notably an alteration in temperature, pressure, or pH, which reduces the substantial amount that can persist dissolved. The notably formed minerals are manganese deposits, phosphorites, glauconites, and carbonates. Hydrogenous sediments are few compared to lithogenous or biogenous sediments, but they have some exciting forms, such as manganese polymetallic nodules[11]. Because any other sediment deposition would instantly encompass the nodules, restricting supplemental nodule growth, they only form in areas with low lithogenous or biogenous sediment accumulation rates [34]. The structure of the nodules varies based on their place and creation circumstances. However, manganese and iron oxides are typically dominant, with minor metallic cobalt, copper, and nickel. They'reThey are only found in the central ocean, far from lithogenous or biogenous sources, but they can build up in large numbers on the seafloor. Also, bacteria are active players in their formation. These are of the prominent marine sediments having great economic value in deep ocean floors, as they contain several commercially valuable metals. The other marine sediments include a good amount of phosphorites, glauconites, and carbonates [1].

Many carbonate masses make the oceans reservoirs of carbonates and a storehouse of carbon dioxide (CO_2). The cosmogenous sediments (CSs) are cosmic in origin. These are microscopic spherules rich in iron and rocky chondrules. Meteoric impacts are the primary reasons for these sediments in oceans. InMetallic elements, such as iridium, in marine sediments, metallic elements, like iridium are indicatives of cosmogenic origin [33].

CSs are obtained from space and have extraterrestrial origins; they are the remnants of the impact of large bodies of space material, such as comets and asteroids [32]. These sediments are rare, but they are mostly found near meteor influence structures or in small amounts combined with a wide range of sediments in all types of marine environments and on land, representing the most insignificantly important one[7]. They form from micrometeorites, which can be silicate, like the mantle, metal, like the core, or a combination of the two. Even though these particles are constantly raining on the earth, CSs are only a minor constituent of ocean sediments[35].



The distribution of the major types of sediments on the ocean floor [1]

Moreover, according to composition, marine sediments are of two main categories; neritic and oceanic. Neritic sediments are deposited near the continents, while the aquatic is the characteristic of deep-ocean basins. The primary outputs of these are the turbidities and glacial and carbonate deposits. Deep ocean basins include sediments such as abyssal clay and calcareous and siliceous oozes. Neritic sediments are mainly lithogenic, but the oceanic six refer to organic and hydrogenic sediments [33].

Furthermore, marine sediments are called pelagic when they accumulate in the deep ocean's abyssal plain, far from terrestrial sources. Pelagic, as a term, refers to the open sea, particularly the upper layers of the ocean away from shore. Slowly accumulating pelagic sediments cover approximately 75% of the deep seafloor [36]. Pelagic sediments; are mainly limited to continental shelves. When mixed with terrigenous, hemipelagic sediments are formed. Aquatic sediments are classified into three types: siliceous and calcareous oozes and red clays[37].

The type of cumulative sediment in a provided area is characterized by its proximity to land, water depth, and overall fertility. Oozes are mainly biogenic, made up of planktonic debris, so although red clays are non-biogenic, with few organic substances. Oozes are not present on the continental margin [37]. The most common pelagic sediment, calcareous ooze, is made up almost entirely of the shells (tests) of foraminifera, coccolithophores, and pteropods and covers 48% of the world ocean's floor area. These oozes are restricted to depths above the carbonate compensation depth (CDD) at the time of burial and accumulate at a rate ranging from 0.3-5 cm/1000 years. Plankton debris with silica shells, including diatoms and radiolaria, comprise siliceous ooze. These oozes are limited to places with high biomass production, including polar oceans and upwelling zones near the equator, and encompass only 15% of the oceanic floor, accumulating at a slightly slower pace (0.2-1 cm/1000 years) than calcareous oozes[38].

. Red or pelagic clays of various compositions accumulate in the ocean's deepest and most remote areas. of the ocean. There is a mixture of fine quartz and clay minerals. They have the colour of oxidised iron minerals, but they can also be brownish. The authigenic deposits formed due to the direct precipitation of seawater and micrometeorites. Its ultimate origin is unknown, but it appears to have come from distant rivers and windblown dust. These deposits cover approximately 38% of the ocean floor and accumulate slower than any other sediment type, averaging only 0.1-0.5 cm/1000 year.

The mechanism of sediment accumulation is another basis of sediments typing. On this, they include: a) Mechanical/clastic sediments; formed as a result of deposition after transportation in water; Sandstone and conglomerates constituted by the erosion of preexisting rocks on the earth's surface or in the oceans, which are then carried to the deposit site by streams, winds, or glaciers. Streams either deposit sediment in floodplains or transport it to the ocean, in which it can establish a delta. At the foot of continental slopes, ocean sediments, particularly turbidites, is deposited [39]. Glaciers also transport sediments that have become frozen within the ice mass. These were thought to be too hostile to sustain life for a long time. Still, they are now identified as distinct biomes governed by microbial populations that maintain active biochemical routes. When glaciers reach the sea, the sediments melt and fall. Glaciers can move very even massive boulders (erratics) much farther than most rivers can. Much smaller particles (sand and dust) are carried far out to sea by the wind [40].

b) Chemical sediments are formed from chemical reactions in seawater that result in the precipitation of water-soluble minute mineral crystals, including limestone, that settles to the seafloor and eventually form a more or less chemically pure layer of sediment. Moreover, evaporation in shallow basins results in a series of evaporite sediments like rock salt and gypsum. [1].

c) Organic sediments formed via accumulation and preservation of OM [39]; as a consequence of plant or animal activity, such as coal and peat constituted by inadequate vegetation decay and subsequent compaction [1]. Continuous sedimentation of organic and inorganic substances occurs in all aquatic ecosystems. Benthic fauna thrives using these resources. Bioturbation, the stirring of sediment by benthic fauna, often destroys the structure of sediments. Recently laid sediments of lakes or seas may contain some detritus food and microorganisms. Benthic fauna feed through the burrowing of these recently laid sediments. The bedding and lamination structures already created in these sediments are disturbed [41]. Deep-ocean sediments, known as pelagic oozes, are formed primarily by the stays of microorganisms (mainly foraminifera and diatoms) from the coastal waters, with minuscule quantities of windblown volcanic and continental dust. In contrast, the aggregation of animal calcite shells formsform limestones.

d) Residual types; formed by weathering, as laterite and bauxite [1].

Importance of Studying and Describing Marine Sediments:

The ocean's bottom is just as essential as the moon's[42]. The oceans contain more than 70% of the earth's crust, with sediments encompassing the vast majority of these floors. As a result, marine soft-sediment communities represent the most diversified single ecotype and are one of the biota's most abundant elements in terms of spatial coverage. They are also essential components of energy flow through food webs, which contributes to the proper functioning of marine ecosystems. They are necessaryThey are important in sedimentary processes like nutrient and carbon cycling, waste breakdown, and elimination [43]. They are also important components of energy flow through food webs, which contributes to the proper functioning of marine ecosystems.

Deep-sea sediments are the most significant global geological deposit, and they serve as documentation of our planet's reaction to events on the surface of the ocean, where the majority of material derives[42]. A better understanding of modern sediment distributions is used to clarify past geologic occurrences, particularly for the last 200 million years of the earth's history, because marine sediments accumulate under particular circumstances (maximum age of the ocean floor). The value of seafloor sediments stems from the fact that they preserve a signature to past climate change. Moreover, exquisitely varved sediments from rapid deposition areas provide great high memories of previous climate variation, and volcanic ash layers have made significant contributions to the study of short-term climate change [44].

Additionally, sediment texture is described to be one of the many possible factors impacting groundfish abundance and distribution [45]. Sediment properties (porosity, permeability, and resistance to displacement); affected by the relative grain size distribution, directly or indirectly impact fish habitat quality determined by growth rates, survival, reproduction, recruitment, and overall abundance [46], and impact the performance of bottom trawls used to survey groundfish stocks [47].

Studying sediment properties is concerned by scholars in hydrology, geomorphology, hydraulics, and env-tal sciences[48]. Particle size is one of the most program management on hydrodynamic circumstances and marine morphology because it has such a significantimpact on sediment transport and ocean ecology [49]. Erosion, deposition, and sorting processes can be investigated by examining the sediment grain-size characteristics of bed materials and suspended sediment, and thus the soil erosion model can be enhanced [50]. Moreover, by analyzing the ancient and modern sedimentary env-ts, It is possible to correlate sediment grain size parameters and the transport process or deposition mechanism[51].

Sediment characteristics are also used to trace sediment sources, which is essential to understanding a catchment's sediment dynamics and sediment budget when designing programs for both on-site soil loss regulation and downstream sediment management [52]. Meanwhile, direct management of sediment sources is critical to reducing uncertainty in the possible performance funding required to sustain and operate viable water supply networks [53].

The results extracted from seafloor specimens are used to 1) investigate past climate changes for environmental forecasting; 2) know the value of benthic habitat on fisheries and another biodiversity; 3) investigate offshore pollution correlations and mechanisms to aid in the maintenance of healthy coasts; 4) collect data on dredged materials for beach replenishment; 5) evaluate the impacts of offshore waste disposal; 6) Learn about and approximate the implications of occurrences like gas hydrate releases that affect slope stability; 7) Locate offshore mineral resources that are strategically important; 8) provide ground-truth values for remotely sensed/satellite data, allowing new methods for environmental estimation and monitoring to be refined; 9) Develop a greater understanding of how the earth and its various environmental systems function; 10) Determine where seabed communications cables, drilling platforms, and other structures will be installed; 11) Improve our knowledge of the role of seafloor sediments in environmental change processes [54].

Fossilized marine creature rock formations make up many tall mountains, such as Mt. Everest. More than half of the exposed rock on the continents is sedimentary rock, which was deposited in ancient oceans and uplifted by plate tectonics. These rocks started as ocean sediments that were lithified and pushed upwards during the formation of mountains [11].

Long cylinders of sediment brought up by ocean floor drilling are known as sediment cores. Oceanographers use these cores to examine the sediment layers accumulated on the seafloor over time. Studying cores as a whole demonstrates information about atmospheric gas composition, past climate, animal and plant evolution, and surface and deep water current movement [55]. A large number of microfossils are also prevalent in deep-sea sediment. The diversification, morphology, and geochemistry of their skeletal remains discovered in ocean sediment cores demonstrated a wealth of information about alterations in oceanic climate and circulation during the Mesozoic and Cenozoic periods. Palaeozoic microfossils such as acritarchs and radiolarians are important [56], larger foraminifera, and ostracodes[57]. Climate and sea-level changes are intimately associated with the revolted different groups, which are punctuated by the effect of mass extinctions and subsequent radiation stages. Because of the keen interest in understanding modifications in the earth's atmosphere that occurred in the past or are happening now, studying marine sediments is an interesting research topic in oceanography [11].

Marine sediment cores are the essential source of data for identifying the characteristics of the seabed, its depositional background, and changes in the environment. They provide raw data for a wide range of research initiatives, including global climate change research findings, slope stability, oil exploration, palaeoceanography, pollution assessment and management, and sea-floor surveys for cable, pipeline, and sea-floor structure placement [58]. Marine sediments are important pollutant sinks and prospective pollutant sources in the marine environment [59].

Marine sediment cores are a major source of information about the long geologic record of long-range-transported (LRT) dust deposition. When compared to other geologic documents of dust input, marine sediments have several advantages [60]; because 1) they are widely and globally dispersed, so eolian sediment records can be obtained anywhere marine cores are elevated; 2) Because they are deposited on the ocean floor, their records are generally free of anomalies (unlike terrestrial records); however, bioturbation can affect them; 3) Foraminifera are commonly found in marine sediment cores, allowing for the extraction of oxygen isotope documentary evidence of glacial-interglacial cycles, which provides a first-order chronology [61]; and 4) Their cores consist of materials such as ostracodes, diatoms, foraminifera, pollen, alkenones, and others, resulting in the development of complementary sets of data that provide independent paleoclimate information[62]. Sediment, in contrast, plays a key role in estuarine physical and biological processes, including shaping the topographic attributes of wetlands, controlling phytoplankton growth, and affecting the transport and fate of water-soluble contaminants through light attenuation [63].

Main marine soil properties:

Soil is one of the last amazing scientific frontiers [64], containing Soil is the earth's top layer and is composed of weathered bedrock (regolith), alive and dead organisms, decaying and residual organic matter (OM), air, minerals, and water [65]. It supports or can support plants out-of-doors and is the literal basis onto which life on Earth is built [66]. Soils are important natural resources to mankind, as an important part of the mountain ecosystem (Liu et al., 2020) and the earth's atmosphere, lithosphere (rocks), biosphere (living matter), atmosphere, and hydrosphere (water); attempting to form the earth's atmosphere, lithosphere (rocks), biosphere (living matter), atmosphere, and hydrosphere (water).

Soil definition depends on who is using the word and the use for which it is put [67]. Soil is merely a bunch of natural bodies found on the planet's surface [68]. It is a biologically active porous medium formed by a series of weathering processes under various influences (biological, topographical, climatic, and geological). It is at the geosphere-biosphere-hydrosphere-atmosphere interface [69]. Soil is now a highly complex dynamic and biogeochemical vital system for diverse land vegetation and soil-inhabiting organisms' life cycles. Soil is a substratum for life on earth, a water and nutrient reservoir, a medium for OM failure, and a participant in numerous biogeochemical cycles **[68]**.

Marine soils, also known as marine deposits, are soil deposits found on the ocean floor. Even though oceans are extremely violent, seabeds are generally calm; as a result, tiny particles would deposit on sea beds. The texture and composition of the water are influenced by its closeness to the land and biological matter. The sediment characteristics, which include physical properties such as grain size distribution and mineral composition, as well as chemical characteristics such as chemical composition and contaminants absorbed by the sediment, are important topics in studying marine systems[70]. Because of geologic and climatic variation over time and distance, soil characteristics vary significantly. Even the most essential property, such as soil thickness, ranges based on the intensity and length of weathering, soil deposition and erosion incidents, and landscape evolution patterns **[68]**. Every soil has three distinct interacting characteristics: physical, biological, and chemical. Knowing the various biological, physical, and chemical processes in the soil can be challenging; nevertheless, knowing soil physical properties can help farmers make better decisions [71].

Soil is one of the world's most valuable natural resources.. Minerals, soil organic matter (SOM), water, and air are present in various proportions. In an ideal soil, air and water fill the pore space and account for about half of the volume, SOM accounts for about 5%, and minerals account for the residual 45-49% [72]**.** The solid fractions of soil, which are usually stable and have a significant impact on other soil physical and chemical properties and processes, such as texture, structure, porosity, and the fraction of pore spaces, are the combination and percentage%age of soil mineral and OM components. These properties, in turn, influence air and water motion in the soil, affecting the soil's capacity perform [73]

The disintegration of different thicknesses of bedrock through physical and/or chemical means, which mantle the ground surface with or without some percentage%age of OM, produces the soil mass of solid particles. It is defined as a mass of solid particles with intervening spaces encompassing either air or water for geologic engineering purposes. The existence and proportion of air and/or water in void spaces, combined with the mineralogical and physical characteristics of solid particles, regulate the use of soil or a fraction of it as a construction material; to support structures or dig up into [74].

The majority of soil physical properties associated with soil organic matter (except clay content) are affected by extreme heat, including soil structure, pore space, and soil aggregation. The following are some of the impacts of excessive soil temperature: 1) Dehydrate clay minerals 2:1 outcomes in less clay and more silt-sized particles in the soil, and 2) heat-induced cracks in sand-sized particles, resulting in sand-sized particle failure and decrease. It also reduces clay content while increasing silt content, 3) decreases water viscosity and soil moisture, enabling more water to percolate through the soil profile; 4) significantly increases evaporation rate and restricts water movement into the soil profile, and 5) promotes soil microbial activity by increasing $CO2$ content in the soil When soil temperature exceeds 30°C, the thermal transformation of iron aluminium oxides, which act as cementing agents for clay particles to form strong silt-sized particles, improves soil aggregate stability [75].

Physical properties of marine soils include a) grain size and sediment texture and structure; b) surface area of soil separates; c) particle density; d) bulk density; e) porosity and pore size; g) colour; h) glossary; and i) soil water [76]. The interactions and interconnections among the constituent parts of the sediment (inorganic solid minerals or biogenous sediment framework and the fluid or gaseous pore filling) that behave as a whole determine the engineering characteristics of marine sediments. Bulk density, porosity, water content, permeability, consolidation, plasticity, and shear strength are all essential engineering characteristics of marine sediments [77].

Shear strength:

Shear strength and consolidation are two of the most important soil properties. They allow for the predictive model of the sea floor's bearing capacity, resistance to penetration, and later settlement under loading over time. It is critical to know these parameters and their variations with burial depth; for the successful implementation of structures on the seafloor or the operation of vehicles and to understand diagenesis by erosion and sediments' redistribution by a gravity-induced mass movement [78].

Grain Size, Sediment Texture and Structure:

Soil particles vary in size, shape, and chemical composition. There are three categories for soil particles/separates by their particle size; sand (the largest; 2-0.05 mm), silt (mid-sized; 0.05-0.022 mm), and clay (the smallest; less than 0.002 mm).

Some are visible with the naked eye, but others are so small that be microscopic, even seen only with an electron microscope, likesuch as clays. Most soils are a mixture of the three, with the relative proportions of each determining the texture of the soil, regardless of chemical or mineralogical composition. Clay-rich soils have a fine texture, whereas sandy soils have a coarse texture. Loam is a textural class that contains about one-fifth clay, with the rest made up equally of sand and silt. The texture is influenced by the chemical and physical properties of the soil. Also, soil texture and structure influence a soil's capacity to transfer and retain water, air, and nutrients by regulating porosity, density, and compactness [79]. Sandy soils have significant pore conditions that enable water to leach, whereas fine-textured soils have a higher capability for water retention. Permeability, infiltration, porosity, shrink-swell rate, and erosion susceptibility are just a few of the properties impacted by soil texture [68].

The water-holding capability of the soil and its cation- and anion-exchange abilities are heavily influenced by its texture. Soil water is essential for microbes because it serves as a solvent for cellular components and a medium for dissolved nutrients to reach their cell surface. Ciliated and flagellated protozoa, Flagellated bacteria, and nematodes require water to move. As a result, soil texture has a direct impact ondirectly impacts bioactivities.

When soil particles and small aggregates clump together, larger aggregates (peds) form, which are the fundamental components of soil, the size, shape, and degree of aggregation (if any) of the structural units (peds/clods), as well as the configuration of the solid soil units into different geometric shapes and the pore spaces between and within the aggregates, are all described by soil structure. Soil structure is amenable. It breaks up quickly when you squeeze good-quality soil because it is friable and has fine aggregates. Coarse, very firm clods or no structure at all characterize poor soil structure [80] Types of soil are classified by their shape, classes by their size, and grades by their stability. According to soil structure, there are four major types of soil structure, these are:

a) Plate-like/Platy (units seen as relatively thin flat horizontal plates or leaflets). When the units/layers are thick, they are referred to as platy, but they are referred to as laminar when they are thin. This is most noteworthy in virgin soils' surface layers, but it can also be found in the subsoil. This type is passed down from the parent material and is influenced by water or ice [81];
b) Prismatic/prism-like (Individual units with vertical faces that range from flat to rounded are noticeably longer vertically, with typically cast or moldy faces of adjoining units, giving a pillar shape). Arid and semiarid regions have subsoil horizons with lengths ranging from 1 to 10 cm. Columnar tops are those with rounded tops [82];
c) Blocky/block-like (Polyhedral equidimensional units with flat or slightly rounded surfaces that are casts of surrounding peds' faces). All three dimensions have the same size. Irregularly six-faced with their three dimensions, more or less equal. When faces are flat and distinct with sharp angular edges, structures are called angular blocky, but when mainly rounded, it is subangular [83].
d) f)); d) wedge (elliptical units with interlocking lenses that terminate in acute angles and are bounded by small slickensides); e) lenticular (spherical or polyhedral units are confined by curved or very irregular faces that are not casts of adjoining peds); f) lenticular (spherical or polyhedral units are confined by curved or very irregular faces that are not casts of adjoining peds (units; prevalently correlated with moist soils, texture classes great in silt or very fine sand and great possible for frost activity, are overlapping lenses parallel to the soil surface and are thickest in the middle and thin towards the edges) [84].

The extent of a soil's micro-and macroaggregates is described by its structure or fabric. A well-aggregated soil is more resistant to rain, and wind erosion has better drainage and is more critical for aerobic population dynamics. More soil is produced for plants and soil biota. Aggregation is a process in which physical and biological features of soil systems become inextricably linked [85].

Poor soil structure limits the depth to which roots can permeate the soil, resulting in poor water and aeration circumstances that limit root growth and limit plants' effective utilization of nutrients and water[86].

Porosity and Permeability:

While porosity is a property reflecting the number and size of pores in the soil, permeability describes the ability of the soil to ease the transport of fluids and their dissolved components[68]. Porosity is determined as the percentage%age of the volume of voids (pore space between grains) to the total sample volume. It is usually expressed as a percentage%age (ratio of non-solid volume to the total soil volume). Because it is a crucial indicator of soil structural quality, assessing the impact of implementing OM on the soil system requires classifying it [87]. It is inversely related to grain size in natural sediments in a complex relationship depending on several factors. Generally, clayey soils have higher porosity and more water than the sandiest. Also, soils finer-textured clayey soils and soils with higher OM content will have higher total porosity than coarse-textured sandy ones and those with less OM [88]. Sand grains and coarse silt-sized particles (single- or mixed-grained) settle speedily to the bottom throughout sedimentation due to gravity and bottom currents. The porosity of naturally deposited sands of uniform size ranges from 23 to about 50%. The porosity of fine silt and clay size, which occupy interstices between grains, decreases[89].

Soil hydraulic characteristics such as hydraulic conductivity, water retention, infiltration, and available water possibility are affected by soil porosity and pore size distribution. Soil pore size influences saturated hydraulic conductivity; coarse-textured soils have larger pores andmore excellent saturated conductivity than fine-textured soils. Porosity and soil water release characteristics directly impact several soil indices, including aeration ability, plant water holding capacity, and relative field capacity. Temperature changes impact on soil porosity and the interface between liquid and solid, especially in heavy loam with high clay content. Furthermore, soil biological action and plant root growth are dependent on soil porosity properties, and warming climate has an effect. Under a warming climate, changes in soil porosity and pore size distribution may influence multiple hydraulic features.

Pore radii, or the spaces between soil particles, range from millimeters between sand grains to micrometers between clay grains. Although sand particles frequently contain quartz and feldspars, silt particles are micaceous, and clayey particles frequently contain layer-type aluminosilicates, soil particles in the three major size groups have varying mineralogical or chemical compositions. Also, significant parameters of soil clay particles are OM and amorphous mineral matter[68].

Soil porosity improves with a more delicate texture, whereas pore size decreases as permeability decreases. Tiny pores prevent the matter from passing through and bring it closer to chemical binding sites on the particle surface, slowing its movement. Clay and humus affect soil porosity and permeability by critical soil grains into aggregates, culminating in a network of larger pores (macropores) that passes water through. The stability of aggregates enhances as humus concentration increases, Especiallyespecially if the hummus comes from grassland vegetation. Pore space and macropore variety are more significant porosity factors than soil texture in soils that have not been significantly disturbed by human activities. Agricultural practices and other human uses of soil decrease average pore size in general.

Permeability is an important engineering property of soil, controlling pore water flow and influencing soil mechanicalbehavior . Variations in soil permeability can have a significant impact on the behavior of a soil structure, such as the failure of an embankment [90]. Because water loss through seepage or infiltration is low in impermeable soil, it is ideal for aquaculture. Permeability varies from one layer to the next because the properties of the soil layers or horizons vary. The soil permeability is determined by the size, texture, structure, and existence of impenetrable layers, such as claypans. For example, permeability is extremely low in clayey soils with platy structures [27].

At the macro scale, soil permeability depicts pore water flow. It's usually measured in centimeters per hour (cm/h), millimeters per hour (mm/h), or centimeters per day (cm/d) or as a permeability coefficient in meters per second (m/s) or centimeters per second (cm/s), or as a permeability coefficient in meters per second (m/s) or centimeters per second (cm/s). Soils appropriate for aquaculture must have a permeability coefficient of less than 5 10^{-6} m/s (pond bottom soils) [90].

Particle Density and Bulk Density:

Density is a function of mass proportion in a fixed quantity or amount of space. The density of a substance concerning the density of the substances around it determines whether it will sink or float, and it is determined by dividing the mass of the matter by its volume [91].

Particle density is the weight of a single soil particle per unit volume, expressed in grammes per cubic centimeter (g/cm^3). It is crucial in understanding and determining other physical properties such as bulk density and porosity. While the average particle density is 2.66 g/cm^3, water has a density of one gramme per cubic centimeter (g/cm^3), and organic matter has a density of 0.8 g/cm^3 [82].

Soil bulk density (BD) is the weight of dry soil (Msolids) split by total soil volume (V_{soil}), which equals the mixed volume of solids and pores, which may encompass air (V_{air}) or water (V_{water}), or both. It is a good measure of soil quality, porosity, productivity, and compaction, which indicate suitability for root growth and soil permeability and are critical for the soil-plant-atmosphere system [92]. Particle density is defined as the weight of a single soil particle per unit volume. Soil porosity reduces as BD rises, and air and water flows are limited. These soils are vulnerable to waterlogging, which leads to anaerobic conditions [93]. Particle density (BD) is the weight of a person soil particle per unit volume. Many factors can influence BD, including 1) soil texture, as BD is dynamic and varies based on edaphic structural conditions. Clays are lighter than silts and sands; 2) types of minerals that make up the soil particles, where some minerals are heavier than others; 3) OM content; as OM has low BD compared to mineral particles; 4) soil compaction that elevates BD; 5) Soil biota, vegetation, and mechanical practices, livestock trampling, agricultural implements, weather and season of the year, and land cover and management are all factors that can influence BD [94].

Generally, uncompacted/light/loose soils (solid rock density~ 2.65 g cm^{-3}), with 50% pore space with a bulk density of 1.1-1.6 g cm^{-3}, are ideal agricultural soils; because of the good aeration and the ease with which roots can penetrate and microbes can navigate. Both texture and OM affect bulk density. As porosity increases, the dry weight decreases for a given soil volume resulting in a lower

Compacted soils have high BD (1.5-1.8 g cm^{-3}), few macropores that restrict root growth and water and air flows. Particle density is the weight of a single soil particle per unit volume; OM weighs little, occupies a large volume, and thus has a lower bulk density. When minerals are dominant in the soil, they have a high particle density (mass/particle volume)and enhance soil bulk density. Soils with a lower bulk density have more plant available water than soils with a high bulkdensity, but they drain rapidly and become drier [95].

The density of seawater is a fundamental oceanographic that functions on temperature, pressure, and salinity. The seawater's average density (ρ) is near 1.025 gm cm^{-3} [91]. The density of seawater is important in the water cycle and climate because it causes ocean currents and circulation, circulates heat, and defines the tendency of seawater to move vertically. If the density of water at the surface is greater than that below, the water will sink to a level equal to its own density. If the density of water at the surface is lower than the density below, water will not sink, and the water column will become constant. In this case, energy input (generally from winds) is required to push water downward. The sinking of surface water occurs most commonly where there is cold air to cool surface waters until they become dense enough to sink thousands of meters. Surface water sinking is a critical mechanism for replenishing deep sea waters. In contrast, the surface water of most oceans (within 50° of the equator) is much warmer and less dense than the cold waters found at depth. Surface waters do not sink under these conditions, so there is no direct contact with deep-sea waters [11].

Because the density of seawater varies with temperature, salinity, and pressure, depth is important. When salt is dissolved in freshwater, the density of the water rises as the water's mass rises. In other words, the higher the concentration of dissolved salts in water, the higher the salinity and density of the water. Water density rises with salinity because solutes have a higher atomic mass than water. Temperature, on the other hand, affects water density. The density of water decreases as it warms. The positive association between pressure and density is well evidenced, indicating that seawater is compressible under the tremendous pressures found in the deep ocean. These characteristics drive the thermohaline circulation of oceanic water. Thermo- refers to density's temperature component, whereas -haline refers to density's salinity element. Water density increases as pressure increases. Furthermore, as pressure increases, water molecules pack together tighter, causing pressure to rise with depth due to the weight of water above. This causes immense modifications in density in seawater with depth, far outnumbering temperature and salinity alterations [96].

Each cubic centimeter of water in the water column would widen if seawater were highly viscous, and density values would be equivalent at all depths. The average sea-level rise would be about 36 meters if the ordinary pressure at 4,000 meters (the rough estimate mean depth of the ocean) were somehow replaced with the average pressure at 2,000 meters, and the area of the oceans stayed unchanged. The temperature of maximum density and the freezing point of water both decrease with the addition of salt, with the temperature of maximum density decreasing faster than the freezing point. At salinities less than 24.7 psu, the maximum density is reached near the ice point, In contrast, greater salt concentrations more typical of the open oceans, the maximum density is never attained naturally. When water is cooled at the surface and density-driven overturn occurs, low-salinity water and freshwater act differently than marine systems because they have the potency to transfer through a density maximum [97].

Surface seawater density varies between 1020 and 1029 kg/m^3, based on the temperature and salinity. The density of seawater is 1023.6 kg/m^3 at 25°C, 35 g/kg salinity, and 1 atm pressure. Seawater with a salinity of 35 contains only 96.5 % water by weight. The residual 3.5 % is made up of dissolved solids. Saltwater has a greater density than freshwater because suspended particles are much heavier than freshwater. Its density varies with salinity and temperature. Seawater has a density of 1026 kg/m^3 at a temperature of 20°C.

Soil Colour:

One of the most useful soil physical features for evaluating and monitoring soil health is soil color [98]. Soil color reflects topography, climatic and environmental circumstances, soil-forming processes and other influences on the soil; as the type of its minerals, soil depth and drainage characters that impact the colour. Except for albedo and the amount of absorbed heat, color has no direct impact on soil behavior. Most soil processes produce colored byproducts that are used as indicators of soil quality and circumstances.

Colour is a soil property that helps to know some of the soil most important characteristics, such as mineral composition, crystal size, shape, and impurities within the crystal structure [99], age and soil processes (chemical alteration, humified OM, carbonate accumulation, etc.). Soil color can determine the depths to wetness or the depth at which water exists in the soil. It denotes the composition of the soil as well as the soil conditions. Water existence in the soil profile affects soil color over long periods caused by changes in the oxidation rate. Similarly, in conjunction with other physical properties, it aids in distinguishing between various types of horizons within the same or different soil profiles. ColorColour also conveys information about the environment. Anaerobic environments, for example, occur when the soil has a high water table or when water settles above an impermeable layer.

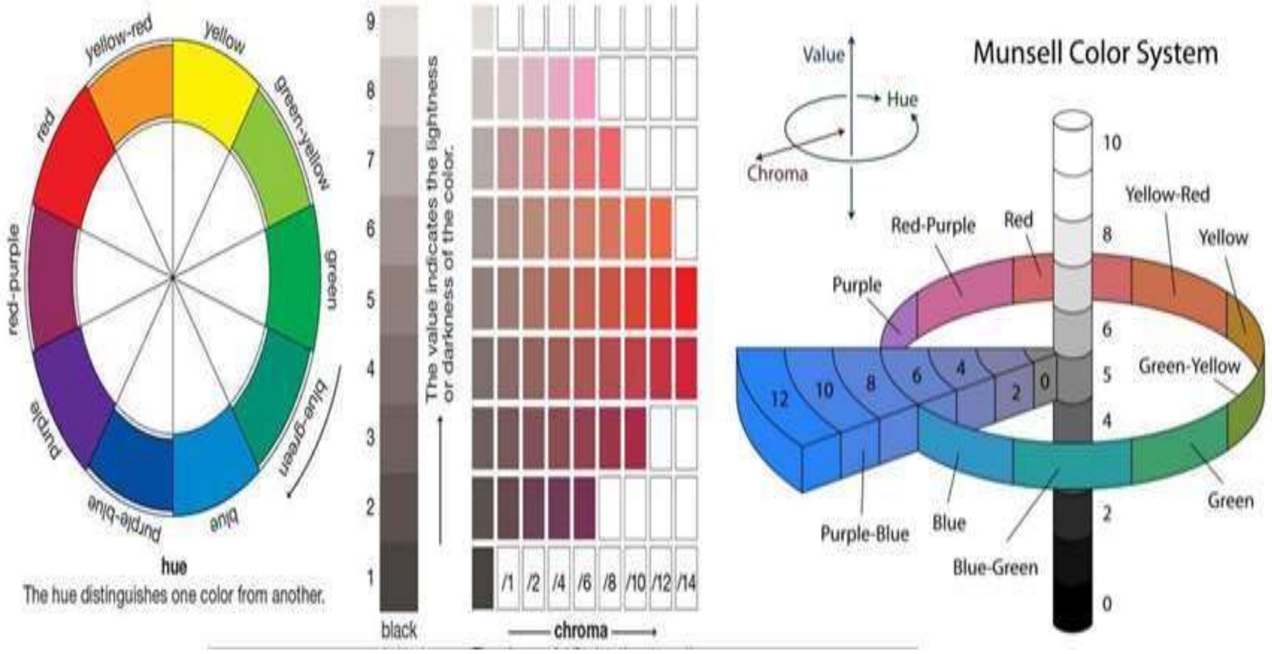
In addition, the soil is made up of a variety of solid particles that are encased in water and/or air. When light strikes a soil particle, some of it is directly reflected (specular reflection), while others are partially transmitted, refracted, absorbed (as heat), and ultimately scattered (reemitted and travel in various directions). Because most soil particles are opaque or translucent, enough light is scattered and diffusely reflected. The diffuse reflection of light after interacting with all particles in its path determines the color of the soil. The final soil's color is an additive function of the color of particles weighted by their %ages. Particle size and arrangement, as well as water content, influence soil color. Smaller soil particles (clay fraction) have a greater impact because 1) they have more surface area for changing light ; 2) they contain the majority of pigmenting substances; 3) their charge and surface area favour physicochemical interactions with colorants,colourants; and 4) scattering decreases as soil particles become coarse. Aggregation, on the other hand, involves: 1) particle configuration and the resulting anisotropic distribution of compounds, which locks some compounds while exposing others to light; 2) increased soil unit size; 3) formation of light-trapping pores,; and 4) when wet, pores filled with water absorb a large amount of light due to increased refractive index.

Soils come in various colors, from grey to black, green to yellowish to red or brown, and vary in purity, lightness or darkness, and intensity. The main determinants of soil color are organic matter (OM) and mineralogy, particularly iron oxides. There were expected to be correlations between soil color and its physical and chemical features [100]. The existence of yellowish particles in the soil, for instance, indicates that more blue (380-480 nm) and green (480-560 nm) light is selectively absorbed while the rest of the light spectrum is diffusely reflected (orange; 590 -630 nm, yellow; 560-590 nm, and red; 630-780 nm) [101]. Gray, white, or colorless salts include silicates, carbonates, sulfates, and other salts. Iron hydroxides, such as hematite, maghemite, goethite, ferrihydrite, and lepidocrocite, are responsible for soil pigmentation. Selective absorptions by electronic transitions in the metal induce their yellowish, brownish, or reddish hues (between adjacent metal ions in different oxidation states or between the metal and ligands). OM is also a common colorant, starting to cause strong absorption in all visible wavelengths and darkening the soil. Less prevalent are theThe Fe (II, III) hydroxy salts are less prevalent, such as jarosite and vivianite, with limited saturation yellow, green, or blue colors, and black monosulfides, pyrite, and Mn oxides [102].

Different elements and their substances have different colors and can influence the overall color of a geological sample when administered in adequate quantities. Vanadium (red/green/purple; depending on implicated compound), chromium (blue-green), manganese (dark red/green/blue); iron (red/blue; varies based on ionic form), cobalt (red/blue); nickel (green/blue/purple), copper (blue/green), zinc (pale blue), and so on are some minerals with accompanying colors (yellow) [100]. Many of these components have different soil colors based on numerous circumstances like redox potential, pH, the existence of OM, and specific clay minerals, all of which strongly impact soil color hue and intensity. These components are available in adequate soil concentrations to produce soil color (usually orange to reddish, though blue in redoximorphic/gleyed soils is common) and the foundation color of mineral particles. Nevertheless, the existence of these components in sufficient quantities behaves as a proxy indicator of specific elemental concentrations.

Minor differences in soil colour can effectively be distinguished with the naked eyes and easily quantified by colour charts; as the Munsell colour chart (MCC) [99]. It is a colorThe colour-designing system is based on a colour-array scheme devised by an American art instructor and painter, Albert H. Munsell, and later). Lately, it has been used as a standard system for determining soil colour and acts as a colorcatalog , colour catalogue, allowing direct comparison of soils from all over the world. Munsell color notations can also describe an archaeological site or compare evidence in a criminal inquiry. Even carpet manufacturers utilize Munsell soil colors to match carpet colors to local soils so that the soil tracked into the house is not visible. The Munsell color tree depicts MCC as a three-dimensional model [103]. The system comprises of three distinct color characteristics that can be depicted as an infrequent color solid in three dimensions as a cylinder. Each color has three characteristics or qualities: the HVC model includes hue (H), value (V), and chroma (C).

While hue refers to the specific color or purity of color; such as red, orange, yellow, green, blue, indigo, violet,..., etc. The measure indicates the lightness or darkness of soil (ranging from black; value 0, at the bottom, to white; value 10, at the top), chroma refers to the brightness or dullness of soil (owing to saturation or brilliance of a color) and light intensity, which ranges from 0 to 10. It is calculated by measuring radially from the center of each slice. A brown soil, for instance, could be described as hue value/chroma (10YR 5/3).



The three-dimensional Munsell color system.

For each of the three color attributes, the system is set up as a numerical scale with visually uniform steps. Each color is logically and visually related to all other colors. A freshly exposed ped of moist soil is used to determine its colour. The moist soil ped is contrasted to the Munsell color chart in direct sunlight. Well-drained soils are brighter than those that are poorly drained. Poorly drained soils produce anaerobic conditions that reduce iron content (Fe^{+3}), culminating in very dull colors. Gleyyed soils have extremely low circumstances and chroma of less than 2 [83]

Soil colour is important for soil classification. It is usually due to 3 main pigments; a) black (from OM), red (from iron and aluminium oxides and white (from silicates and salts). In general, it indicates the status of OM, water (moisture) content and degree of saturation, drainage, mineral composition of the soil (derived from the constituents of the parent materials); as manganese and dark minerals, redox states, and nature and abundance of iron [104]. The color of humus-rich soils ranges from dark grey to black. Darker colors result from soil saturation with moisture. The color of well-drained soils ranges from yellowish-red to brown. Soils have a reddish color due to oxidized iron compounds. Due to obstructed drainage (gleying), iron compounds produce green colors in subsoils due to blocked drainage (gleying). When present in high concentrations, Manganese substances can cause a black color. Colorcolour. Colour, on the other hand, does not affect soil behavior.

Soil can be mottle or have the floor (matrix) color (discontinuous spotted streaks and small spots or blotches of contrasting color, usually red or yellow), which are different. The degree of difference between matrix and mottled patterns lies in three areas of contrast' a) faint (difficult to see in place, b) distinct (easy to see in the test pit) and c) prominent (easy to see from outside the pit). The aeration or drainage of the soil is linked to both matrix and mottle colors. Low-chroma matrix colors with or without mottles are called gleyed colors (greenish).If the soil is gleyed, it will most likely be dry for most of the year (due to postponed moisture saturation) . Mottles appear in soils that undergo seasonal reduction and oxidation due to changing groundwater levels.

Soil temperature:

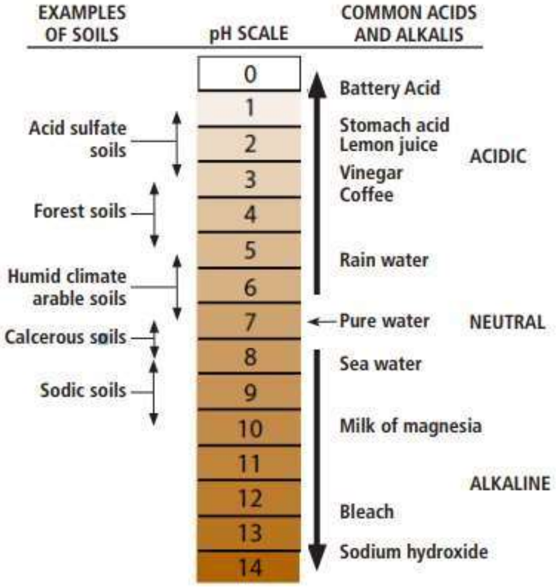
Soil temperature (ST) is a significant major limitation, particularly in the tropics. STs approximate air temperatures at about 50 cm depth. Soil temperatures can be limiting in two ways: incredibly high temperatures in certain sandy top soils and low temperatures in the tropical highlands. The temperature of the air and the temperature of the soil are inextricably linked. Warmer air temperatures raise soil temperatures and hasten the rate of biogeochemical processes. Soil temperature is governed by radiation gains and losses at the surface, heat conduction through the soil profile, evaporation, and convective transfer through the motion of gas and water. Soil temperature, like soil moisture, is a primary mover in most soil processes. Warmer soil temperatures hasten soil processes such as OM decomposition, increased microbiological activity, faster nutrient release, increased nitrification rate, and overall chemical weathering of minerals. Nevertheless, soil temperatures will be influenced by the type of vegetation growing on its surface, which may change as a consequence of climate change or adaptation management.

Textural changes in the soil profile are caused by temperature changes over time, with potential ramifications for the biogeochemical cycling of vital elements like carbon (C), phosphorus (P), and nitrogen (N). Temperature-mediated changes in vegetation dynamics, such as above- and belowground plant productivity, microbial biomass activity,community structure, and OM decomposition, cause soil structure to alter more quickly. Reduced aboveground plant productivity and root biomass can also contribute to higher soil erosion rates, resulting in a loss of soil structural stability. This structural change is most likely to occur in transitional areas with changing vegetation. Because biochemical by-products of microbial decomposition are major sources of stable SOM, which is guarded within soil aggregates and through mineral-SOM relationships, the impacts of change in temperature on soil biota dynamics are significant for soil structural dynamics. Warming causes a reduction in microbial biomass and correlated products, which has a negative impact on soil structure. Under warmer conditions, temperature-mediated changes in soil structure that cause waterlogging (due to a reduction in snow cover and/or thaw) and salinity/alkalinity have a negative impact on soil biota and soil structure. Soil aggregates affect the spatial distribution of soil microorganisms, carbon, and nutrient (N and P) cycling, as well as water and oxygen fluxes, causing different biogeochemical responses to soil warming by restricting their movements or interactions in soil. Due to SOM's elevated physical protection in micro aggregates, both soil microbial community composition and soil enzyme activity were found to be more significant to soil warming in macroaggregates than in micro aggregates.

Soil warming could wreak havoc on aggregate stability and promote C loss. Warmer soils may result in faster macroaggregate turnover, owing to quicker OM decomposition. The malfunction of macroaggregates into micro aggregates and silt-clay fractions due to loss of soil structural stability and an increased risk of physically guarded SOC damage is one of the possible ramifications of faster macro aggregate turnover [105]. Soil warming may have a negative impact on soil aggregate stability and promote C. Soil warming causes the loss of chemically guarded SOC, which destabilizes the large stock of C that is chemically stabilized in soil. Temperature and density also have an inverse relationship. As the temperature rises, the space between water molecules expands, resulting in a decrease in water density.

Soil pH:

Naturally, soils can be neutral, acidic, or alkaline, as demonstrated by the pH value, which measures hydrogen ion concentration. The proper pH is essential for healthy plant growth. Soil pH is affected by the long-term impacts of multiple soil management practices. The pH of most soils is between 3.5 and 10. Soil pH in more significantrainfall areas generally ranges from 5 to 7, whereas it ranges from 6.5 to 9 in drierregions. Acid sulphatesoils are highly acidic (pH less than 4).



The soil pH scale

Natural soil pH depends on parent materials and the weathering processes acting on them, such as climate, vegetation, topography, and time, which cause a lowering of pH (increase in acidity) over time. It has enormous influences on soil biogeochemical processes, so characterized as the master soil variable that affects a plethora of soil biological, chemical, physical features, and functions that influence plant growth and biomass yield

Some biogeochemical processes and their relations with soil pH

The determination of pH aids in the understanding of many soil reactions. The pH differences between soils have been linked to a variety of other soil parameters. The number of soluble chemicals in soil water, and thus the amount of nutrients available to plants, is affected by soil pH. Some nutrients are more available to plants in acidic conditions, while others are more available in alkaline conditions, and most mineral nutrients are readily accessible for the plant in neutral conditions. The formation of highly acidic soils ($pH < 5.5$); due to toxicity with aluminium and manganese, deficiency of magnesium and/or calcium and low standards of essential plant nutrients; as phosphorus and molybdenum, results in poor plant growth. Alkaline soils have nutrient deficiencies such as zinc, copper, boron, and manganese. Soils with extreme soil alkalinity ($pH > 9$) have high sodium levels. The ideal soil pH range is between 5.5 and 7.5, so make every effort to check soil pH levels regularly. Early detection of soil pH issues is critical, as long-term impairments can be costly and difficult to correct. High OM and clay content soils are more resistant to pH changes and necessitate higher application rates.

The pH of marine waters is nearer to 8.2, while the pH of most natural freshwaters ranges from 6.5 to 8.0 . Water's or solution alkalinity is tits quantitative capacity to buffer or neutralize an acid, thus used to measure water'sas a measurement of water's ability to resist changes in pH. Water, in general, has some resistance to pH change due to the impact of the carbonate-buffer system, in which hydroxyl ions produced during bicarbonate hydrolysis neutralize H+ ions and keep pH near consistent. Weathering of silicate or carbonate minerals produces bicarbonate ions (HCO^{3-}) as rainwater passes through the soil zone [106].

The pH changes due to changes in dissolved carbon dioxide concentrations, alkalinity, and temperature (to a lesser extent). Because different ions are implicated in acid-base reactions and salt concentrations impact various equilibrium constants, the magnitude of alteration varies with salinity. In natural waters, pH rises with salinity until it reaches saturation with calcium carbonates ($CaCO_3$). When $CaCO_3$ precipitates, the carbonate-alkalinity of water decreases, resulting in a reduction in water's buffering capacity and a reduction in pH [107].

Water is considered healthy when the pH range is 7.5–8.5. The pH of most aquatic organisms and some bacterial processes must be within a certain range. Most aquatic creatures prefer a range of 6.5-9.0; outside of which optimal ranges, organisms are stressed or die. The more sensitive a species, the more affected by pH changes. The growth of fish is so good in the range of 7-8. When the pH of an organism, such as microbes, changes above or below its desired range, physiological processes are harmed. When the pH of the water is too low for fish, it can cause physical harm to the gills, skin, and eyes, and skin damage enhances susceptibility to fungal infections. Under eutrophic conditions, pH is driven to greater extremes more frequently, enabling algal species with tolerance to extreme pH levels to grow and dominate communities, possibly forming algal blooms. Furthermore, pH changes indirectly affect aquatic organisms and can change the biological accessibility of metals, nutrient speciation, and the toxicities of ammonium, aluminium, and cyanide. Heavy metal accumulation in organism tissues causes chronic illness and potential harm to the human population [108]. Increased pH values make the binding electrostatic forces between viruses and particles to be overcome; as a result, their release into the water column is facilitated. Furthermore, pH affects $CaCO_3$ solubility, which is essential for shell-forming organisms. If the water is too acidic, shell growth and calcification are impeded.

Water Salinity and Electric Conductivity:

Everyone recognizes that seawater is salty, but few realize that even minor changes in ocean surface salinity can significantly impact the water cycle and ocean circulation. . Definite processes have caused the ocean to become salty throughout Earth's history. Weathering of rocks releases minerals into the ocean, such as salts, and ocean water evaporation and sea ice formation boost oceanic salinity. Processes that cause salinity to decrease, including the continuous freshwater input from rivers, precipitation of rain and snow, and ice melting, balance these salinity-increasing factors.

Few people realize that even small changes in ocean surface salinity can significantly affect the water cycle and ocean circulation. Ocean surface salinity is crucial in understanding how freshwater input and output impact ocean dynamics because the ocean accounts for 86 percent% of global evaporation and 78 percent% of global precipitation. The driving factors of climate and ocean circulation are changes in seawater density, which is determined by temperature and salinity [109]. In some areas, the surface water becomes cool and saline, as well as dense enough to sink to enormous depths. Because the oceans store more heat in their upper three meters than the entire atmosphere, density-controlled circulation is essential for transporting heat and maintaining the Earth's climate [110].

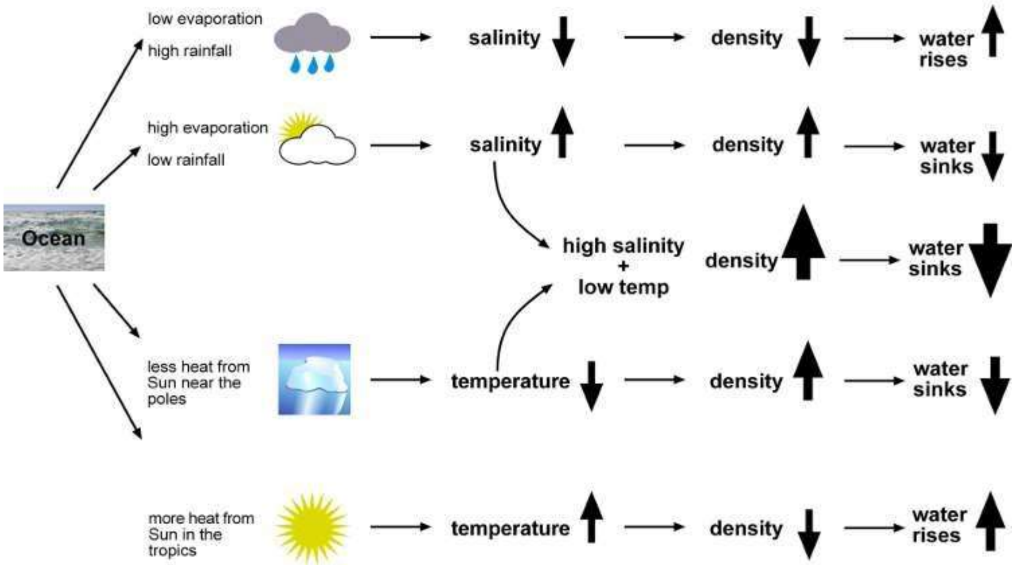
Moreover, salinity is significant in the global water cycle, as changes in the collective evaporation and precipitation contexts are mirrored in sea surface salinity (SSS). Subtropical oceans, such as the Red Sea, have high evaporation and impact water salinity, whereas high precipitation regions have low surface salinities. Dilution reduces salinity near-shore regions near large freshwater sources. Such water changes have an effect not only on water circulation but also on the climate Sommer et al., 2015 River runoff, which outcomes from the weathering of continents over millions of years, is the primary source of salts in the ocean, and the dissolved components become fairly spread in the ocean as an outcome of combining. The total concentration of soluble salts varies significantly from one location to the next. Evaporation and dilution by freshwater from rain and river runoff cause these variations. Only at the sea surface do evaporation and dilution occur[47].

Salinity is an important determinant of the mixing system in the water body; because density variation is correlated with salinity variation. Salinity stratify-cation inhibits vertical water mixing, which is a crucial process in remobilizing OM from the seabed to the water column [111]. As per the International Council for the Exploration of the Sea, absolute or ideal salinity is the total sum of dissolved material in grams in one kilogram of seawater after all carbonate has been converted to oxide, bromine, and iodine have been supplemented by chlorine, and all organic matter has been entirely oxidized [112]. Practically, Salinity is measured in psu (practical salinity unit) or ppt (parts per thousand), which is very near to psu [113]. The salinity range in the open ocean is from 33-37 gm salt/kg seawater or psu, with a mean value approximates 35 psu; due to processes at Earth’s surface that add or remove freshwater. Salinity is expressed as S ¼ 35.00 or S ¼ 35.00 psu. The practical salinity scale 1978 is the name given to the algorithm that is broadly used to determine salinity from conductivity and temperature (PSS 78).

In addition, salinity is a vital ecological parameter; due to its role in certain chemical processes in the marine env-t. Most aquatic organisms function optimally within a narrow range of salinity, and when salinity changes outside of this range, an organism's ability to regulate internal ion concentration may be lost. As a result, osmoregulation becomes so costly in energy that the organisms succumb to biotic pressures such as competition, disease, predation, or parasitism. Any drastic changes in the salinity of coastal water alter the distributions of macrobenthos, sessile organisms and rooted vegetation.

Salinity distributions in coastal waterways are divided into four categories: stratified, partly combined, fully combined, and inverse. Stratified coastal waterways are distinguished by a distinct rise in salinity with increasing water depth in stratified circumstances. When riverine flow is sufficient to create a plume of low-density freshwater (1000 kg m^{-3} at 20^oC) flowing over elevated-density seawater (1025 kg m^{-3} at 20^oC), and tidal currents and waves are inadequate to mix the water column, and this phenomenon known as stratification occurs. In shallow water, bottom waters are cut off from dissolved oxygen-enhancement processes (gas exchange across the photosynthesis and water surface), resulting in anoxic and hypoxic conditions [114]. Tides that create turbulence are partially mixed coastal waterways, which promotes vertical mixing. However, tidal currents are insufficiently strong to fully combine the water column, so salinity varies vertically and horizontally [115]. Fully- or well-mixed conditions occur in coastal waterways where the energy of the tides, rivers, or waves creates enough turbulence to mix the water column, resulting in a constant fluid density over the depth and a varying density in the longitudinal direction from that of seawater (1026 kg/m^3); thus, salinity is uniform throughout the column but varies between the riverine and oceanic ends of the estuary. Inverse estuaries form when evaporation rates exceed freshwater inflow rates, resulting in hypersaline conditions. Over the top of this layer, less dense (lower salinity) seawater flows into the estuary from the sea or ocean. Depending on the time of year and the location of the estuary, an estuary may fall into any of these salinity zones.

Seawater is a complicated mixture that contains nearly all of the elements known to science. Some of the more abundant elements as a %age of the total mass of dissolved material are sulphate ion (7.7%), sodium ion (30.7%), chlorine ion (55.0%), magnesium ion (3.6%), calcium ion (1.2%), and potassium ion (1.1%). While the total concentration of dissolved matter varies by location, the ratios of the more abundant element on earth remain stable [91].



The soil pH scale

Salinity definitions based on chemical observations have recently been supplemented by one based on the electrical conductivity of seawater, which varies with salinity and temperature. Despite the fact thatEven though the preferred international convention is to useuses no units for salinity, this conductivity-based quantity is recognized as functional salinity, and the symbol psu sometimes symbolizes it for practical salinity units. The temperature significantly impacts electrical conductivity, but ion material and salinity have a minor residual effect.. As a result, in order to determine the practical salinity, the temperature must be precisely controlled or analyzed during the measurement technique [116].

The capacity of an aqueous solution to carry an electric current is known as conductivity. Another type of field observation can be used to determine whether or not a groundwater well is stable. The concentration of obtainable charge carriers (ionic), the valence state of the ionized species in solution, their mobility, and the temperature at which the decision is made all influence it [117].

A water sample's conductivity is a good indicator of the levels of dissolved salts, bases, and acids. Organic compounds (those that do not dissolve in water) have low conductivity, whereas inorganic compounds do. The speed with which different sizes and charge ions move in water is affected by temperature because higher temperatures make water less viscous, leading to changes in conductivity. The salt material of soil water extract is measured by electrical conductivity, which is a proxy for salinity [11].

Conductivity is a significant property of seawater that improves almost linearly with temperature and slightly less linearly with concentration and is measured in Siemens per meters (S/m). Conductivity, frequently used to ascertain the salinity of seawater, gives information about the solvent structure. Ionic conductivity is related to an ion's electrical mobility in solution. The conductance of a solution is inversely proportional to the resistance it provides to the flow of current. The mechanism of the conduction process in aqueous electrolyte solutions is not well understood, but ions in a solution or seawater are constantly random walking, and the superposition of an electric field induces them to walk in the direction dictated by the area. Ions, on the other hand, are hydrated and relatively bulky, and the structure of water itself is most likely discontinuous. The conductivity of an aqueous electrolyte solution increases almost linearly with the temperature at one atmosphere and over a temperature range of 0 to 30°C. Increasing the temperature, for example, enhances the size and number of vacancies into which an ion can jump and the energy available to the ion to make the jump.. Individual ion conductance cannot be calculated without making a non-thermodynamic assumption. It has been demonstrated that more firmly hydrated ions face greater resistance to their motion through the solution [118].

Conclusions:

The study of the physical and chemical properties of aquatic soils mainly aims to complement any study in those water bodies because of their direct and indirect impact on the rest of the vital and non-vital data of those environments as they interact with each other affect each other negatively and positively. Temperature, salinity, pH, and even depths, soil texture, and colorcolour affect the rest of the factors and thus influence the presence or loss of some living organisms and the study of the area's biodiversity. Chemical elements and components play a major role in this as well. The percentages%ages of some heavy metals may be high or rare, and even essential ones are deadly and reach the point of toxicity. Population activity, climatic changes, weather erosion factors, pollutants, human waste, urban progress, and other aspects of intervention have contributed to the disruption of these natural components and their contents of living and non-living elements

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