Explain the processes which may occur during the interval between the moment of evaporation and all forms of precipitation at Earth's surface.

Earth's outer fluid spheres operates as an interconnected system of moving air and water that fundamentally creates and controls the hydrological system and global climate. The processes of evaporation and precipitation are deeply embedded within this system of dynamic moisture, energy and heat movement and recycling. Indeed, it is the distribution of atmospheric moisture 'sources' (P < E) within the tropics and 'sinks' (P > E) within the midlatitudes that facilitate meridional heat transfer and the regulation of global temperature to habitable levels. This distinct spatiality between source and sink occurs because evaporation rate increases with temperature and warm air holds more water vapour than cold air. Consequently, evaporation occurs largely within the high-insolation tropics; the Hadley cell, its most observable manifestation. In order to explore the processes that occur during the interval between evaporation and precipitation, this paper follows the logical and temporal progression provided by the rising parcel itself. Primarily, I explore the mechanical processes of surface evaporation and vertical motion that supply moisture to the troposphere. It is these vertical movements of the evaporated air parcel that facilitate condensation and the formation of precipitation clouds in the upper troposphere; in essence carrying moisture from sites of evaporation - the warm ocean or tropical rainforest - to sites of condensation.

I explore how such vertical mobilisation is the result of temperature and pressure differentials between the warm, moisture parcel and its surrounding cooler air. Furthermore, by observing the interactions of adiabatic, saturated and environmental lapse rates, particularly at the condensation level, I suggest that instability - induced by the condensing, vapour particles releasing latent heat - creates the optimal conditions for precipitation. Within the cloud structure, I explore the microphysical processes of collision, coalescence and subsequent supersaturation that occur to form large enough droplets for precipitation. These droplets - rain, snow, hail, graupel, sleet, dew and rime - differ widely in internal properties because of the different conditions under which they formed. Finally, I suggest that these micro-chemical processes of condensation are deeply intertwined with the macro-mechanical processes of moving air. With reference to thunderstorms, it is shown that the interactions of air turbulence dynamics and microphysical cloud chemistry often serve to create self-perpetuating, convective weather systems. The longevity of these meso-scale weather systems - such as hurricanes - is facilitated by the constant undercutting and overcutting of warm and cooler air parcels. Such fluid behaviours and their associated positive feedbacks are fundamentally influenced by the heat and energy fluxes of the evaporation-precipitation cycle.

The process of evaporation acts as the primary stage in the formation of precipitation. Zones of intense evaporation tend to cluster geographically; in the tropical oceans, associated with Trade Wind belts, and over equatorial land areas in response to high solar radiation receipts and luxuriant vegetation growth. (Barry & Chorley, 2009). These distinct spatialities of evaporation, predominantly in the tropics are due to preferential availabilities of solar energy, surface-air difference in vapour pressure and wind turbidity. Evaporation requires an energy source at a surface that is supplied with moisture – solar energy is required to break the strong, covalent intermolecular attractions within the liquid water molecules. As this occurs the surrounding surface drops in temperature due to the latent heat of vaporisation wherein energy is absorbed into the water molecules. The latent heat of vaporisation required to evaporate 1kg of water at 0°C is 2.5 x 10^6 J. This same latent heat is significant in later condensational processes, energising parcels of air to warm and behave more unstably. These associated releases and absorption of heat in humid air are what regulate the diurnal ranges of temperature, such that evaporation taking place during the hot day absorbs heat - a cooling process - and condensation at colder nights releases heat - a warming process.

The remaining major determinants of evaporation - surface-air differences in vapour pressure and turbulent winds - are closely interlinked. Vapour pressure refers to the fact that at any given temperature, there is a limit to the density of water vapour in the air, such that often when air is enclosed above an evaporating water surface, an equilibrium is reached between the water molecules evaporating and those returning to the reservoir. At this stage the air has reached saturation, and the corresponding pressure is termed the saturation vapour pressure. The Clausius-Clapeyron relationship observes how this saturation vapour pressure increases with temperature, such that 'at 20°C air may hold 15g of water, but at -10°C it only holds a mere 1.5g' (O'Hare, 2005). That, at -10°C, less moisture can be held becomes significant in later microphysical processes shaping precipitation. Once the rising parcel of air has cooled to these temperatures, vapour beyond the 1.5g limit will begin to condense. Returning to ground level however, it is the presence of turbulent surface air that deters surface air from reaching this saturation vapour pressure, even if air temperature and thus moisture capacity is relatively low.

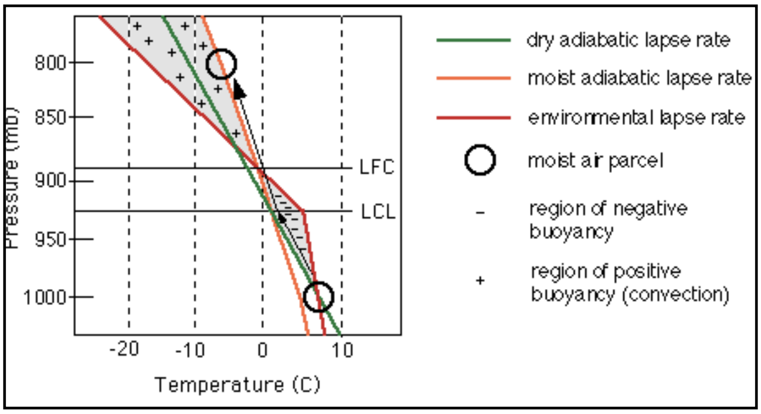
This is because air is continuously replaced with drier air, thus reducing the potential for water molecules to saturate it. Thus under hot, humid weather conditions with strong winds, rates of evaporation are strongly enhanced. Moisture content in the Inter-Tropical convergence zone, as such, is relatively high due to intensely localised evaporation facilitated by strong solar heating, high surface temperatures and strong mixing winds. Similarly, the Doldrums create little evaporation because of its low winds and static air, and the subsequent high proximity of the vapour pressure in the air to be near its saturation level. Nevertheless, 87% of atmospheric moisture is supplied over the ocean surfaces that hold an unlimited water supply. Continents provide the remaining 13%, with evaporation activity concentrated in the tropical rainforest regions. Transpiration occurs in these vastly vegetated regions when the vapour pressure in leaf cells is greater than the atmospheric vapour pressure. Such a pressure gradient results in moisture loss via osmosis to the surrounding air. As Barry posits, this process is significant inter-continentally, '52% of annual evaporation being due to transpiration, with 28% down to soil evaporation and 20% a result of interception' (Barry, 2010) Manufactured through mechanisms of solar heating, wind turbidity and vapour pressure differentials, evaporation serves to create the rising air parcels of moisture necessary for precipitation.

Atmospheric gases, including water vapour, obey Boyle's law, in relation to changes in pressure and temperature. Boyle's law, states that, at a constant temperature, the volume (V) of a mass of gas varies inversely with its pressure (P): $p \propto 1/v \propto d$ (Where p = pressure, d = density and v = volume). Thus as pressure decreases, volume increases; a relationship that underlies much of the vertical motion of air parcels. Within the atmosphere, this pressure decrease with height results because the amount of overlying gas decreases with height. Consequentially, a rising parcel of air that experiences less pressure will expand and cool. The parcel cools because its energy is exerted on its surroundings through the physical work of expansion at the expense of internal energy. The parcel does not however interact and impart energy with its surroundings; it is thus adiabatic because the falling temperature is purely that of internal metamorphosis. In such an occurrence, the temperature of the parcel falls at the constant dry adiabatic lapse rate (DALR) of 9.8°C/km. Prolonged cooling of the air parcel at such a rate invariably produces condensation, and when this happens latent heat is liberated, counteracting the dry adiabatic temperature decrease. Therefore, rising and saturated air, energised and heated by the latent heat release of suspended moisture, cools at a slower rate - the saturated adiabatic lapse rate (SALR) - than air that is unsaturated. Unlike the linear DALR however, the SALR is dynamic and varies with temperature because air at higher temperatures is able to hold more moisture and therefore on condensation release a greater quantity of latent heat. Thus the SALR at high temperatures may be as low as 4°C/km with this rate increasing as temperatures fall to 9°C/km. This close relationship of moisture and temperature equates to the SALR being non-reversible, in a descending cloud air saturation cannot persist because droplets evaporate. Whilst the DALR and SALR refer to individual parcels of buoyant, rising air, the ELR refers to the vertical profile of surrounding, atmospheric air and thus to the 'actual temperature decrease with height on any occasion' (Barry & Chorley, 2010).

Adiabatic changes tend to operate in the atmosphere because air is fundamentally a poor thermal conductor, thus there is little interaction between the air parcel, its thermal identity and surrounding air. However near the Earth's surface, most temperature changes are non-adiabatic because of energy transfer from the surface and the tendency of air to mix and modify its characteristics by lateral movement and turbulence. These vertical non-adiabatic and adiabatic motions of air parcels may originate from multiple source events: gradual uplift of air over a wide area in association with a low pressure system; thermal convection on the local cumulus scale; uplift by mechanical turbulence (forced convection) or ascent over an orographic barrier. In forced convection uplift is produced by mechanical forces such as flow over orographic barriers or turbulence due to surface friction and the convergence of air. However, in natural or free convection, movement is brought about through the buoyancy effects of rising convective cells. In actuality, both forced and free convection often occur together as mixed convection, wind effects being significant and free convection acting as a modifying influence.

Processes of evaporation and vertical motion along a temperature gradient (defined by lapse rates) feed into systems of atmospheric stability and instability. Movement is created because of the interactions of differential lapse rates between individual parcels of air and surrounding air. If the air into which a thermal rises is warm or has a low lapse rate, a thermal rising through it will soon stall. These conditions are said to be stable because the rising parcel is cooler and more dense than its surroundings and therefore tends to revert to its former level. If, however, the surrounding air is relatively cooler than the parcel, instability is said to exist because the thermal has a tendency to move away from its original level once set in motion. Instability is often facilitated by horizontal motions of air. In the Midwest USA, air with a marginally stable lapse rate, originating over high plains, may be undercut by warm, moist air from the Gulf of Mexico to form instability. Similarly, atmospheric instability is created when air above the parcel cools, a common occurrence behind Atlantic cold fronts when the horizontal motions of cold air over warm water create steep temperature differentials. This fine balance between atmospheric stability and instability is embedded in a knowledge of lapse rates.

Stable air exists when the environment lapse rate lies to the right of the DALR and SALR; surrounding air is warmer than the thermal and thus suppresses any buoyant vertical motion. However, if local surface heating causes the environmental lapse rate near the surface to exceed the dry adiabatic lapse rate (surrounding air losing heat more rapidly than the thermal), then the adiabatic cooling of a convective air parcels allows it to remain warmer and less dense than the surrounding air, so it continue to rise through buoyancy. Barry et. al categorise these thermal regimes of stability



and instability that determine the ability of air at rest to remain laminar or become turbulent through buoyancy: Absolutely Stable (ELR < SALR); Saturated Neutral (ELR = SALR); Conditionally Unstable (SALR < ELR < DALR); Dry Neutral (ELR = DALR); Absolutely Unstable (ELR > DALR).

Whilst theoretically, absolute stability is possible, in reality much of the Earth's atmosphere is often in a state of conditional instability. In this situation, the atmosphere is layered, being stable for air which has not reached saturation point, but unstable for saturated air. The fluid boundary between stable and unstable air as such is located where condensation occurs; at sites of highly energised, latent heat release. The presence of water vapour is thus key in the creation of instability, such as that seen in hurricanes. Reiterating Boyle's law, rising air has a tendency to cool because of energy loss in expansion. Such progressive cooling loosens the air parcel's capacity to retain moisture. As the air cools, the evaporation rate decreases more rapidly than the condensation rate and eventually the air becomes saturated, no more moisture can be held. It is said to have reached the dew point, the temperature at which atmospheric saturation arises following cooling without a change in moisture levels. After reaching this point of 100% relative humidity, condensation occurs with clouds forming above the condensation level. This height is called the lifting condensation level (LCL) wherein Tair = Tdew; 'lifting' as such because it serves to energise and heat the parcel, through the release of latent heat, into further vertical expansion. As water changes from its vapour state to a liquid, latent heat is liberated, counteracting the dry adiabatic cooling of ascent. Latent heating thus skewers the lapse rate, providing the parcel with positive buoyancy relative to the ambient air. As such it will continue to ascend and even accelerate upwards in turbulent manner. The buoyant energy made available by lifting a saturated air parcel to its level of equilibrium is known as convective available potential energy. Instability, induced by latent heat release, predominates in the tropics because warm air is able to hold a lot of moisture, and thus, on cooling release a lot of latent heat. Tropical depressions amplify into cyclones by such energy release and the strong convection it creates. The updraughts are strong, facilitating the rapid and violent cooling of air and condensation of water vapour.

Within the turbulent updraughts of these storms, the condensational latent heat released from rapid droplet growth enhances upward motion; whilst in the downdrafts, evaporation in the subsaturated environment strengthens the downward motion. These downward motions become significant in convectional precipitation events, aiding the collision of ice crystals in lower layers. Cold air conversely is able to hold far less moisture, so the heat production, and creation of turbulence during condensation is much less. (Addison et. al, 2008) Furthermore both the convective and lifting condensation levels behave fluidly; as surface temperature increases with little change of dew-point, they rise. Once unstable air is lifted it continues to rise, until it reaches a layer of stable air such as the tropopause, temperature relationships are inverted and increase rather than decrease with height. The parcel has the capacity to penetrate this inversion layer because it contains momentum. It is now, however, cooler than the surroundings and will sink back down and come to a rest at some equilibrium level. Instability between the lifting condensation level and the inversion layer thus dominate the processes of cloud formation and precipitation. As such cloud-base height is the height at which the dew point is reached, whilst the overall height of a cloud is dependent on the stability of the atmosphere, or, whether the air parcels reach the no man's land inversion layer of the tropopause.

Whilst the formation of clouds depends largely on atmospheric instability and vertical motion, it too involves microscale processes. The interactions of water vapour and condensation nuclei underlie much of a cloud's lifespan. Usually condensation occurs on a foreign surface' such is the case on morning window panes. In the atmosphere however condensation begins on hygroscopic nuclei. These are microscopic particles - aerosols - the surface of which have the property of wettability. These dusts, smokes, salts and chemical compounds suspended in the troposphere, originate oceanically, from the bursting air bubbles in foam. On landmass, the rainforest of the tropics produce aerosols through biomass burning. Cities are a further source, producing condensation nuclei by 'the conversion of atmospheric trace gas to particles through photochemical reactions'(Barry & Chorley, 2010). These nuclei range in size from 0.001um radius, which are ineffective owing to the high supersaturation required for their activation to giants of over 10um, which do not remain airborne for very long. Condensing water droplets form around these soluble nuclei. In the primary stages of droplet growth, the solution effect is predominant and small drops grow more quickly than large ones. However as the size of a droplet increases, its growth rate by condensation decreases. As O'Hare posits, 'radial growth slows down as the drop increases because there is a greater surface area to cover with every increment of radius.' (O'Hare, 2005) These condensation rates are further complicated and limited by the speed with which the release latent heat can be lost from the drop by conduction to the air; this heat reduces the vapour gradient. That rain clouds may form within an hour appears to be dissonant from this theory of slow growth by condensation. However,   it does not disprove the process, it simply reveals that there are other, more subtle ones coexisting to form precipitation.

The Bergeron-Findeisen theory of water droplet growth is based on the fact that at subzero temperatures the atmospheric vapour pressure decreases more rapidly over ice surface than over water (Barry & Chorley, 2010). As such the saturation vapour pressure over water becomes greater than over ice, a process particularly recurrent between -5°C and -25°C, where the difference exceeds 0.2mb. Furthermore it is within these subzero temperatures that water droplets become supercooled, the surface tension of the water allowing them to stay liquid well below the normal freezing point. The differences in saturation vapour pressure are experienced acutely when ice crystals and supercooled water droplets coexist in a cloud. The interactions involved are often turbulent, as Dhudia writes, 'drops tend to evaporate and direct deposition takes place from the vapour on the ice crystals' (Dhudia, 1996). These interactions are catalysed by the fact water vapour is supersaturated. Thus rather than being saturated, like water vapour interacting with water droplets at 100% relative humidity; that same of water vapour is supersaturated because it is now interacting with foreign, ice droplets. It is such that 'at -10°C, air saturated with respect to liquid water is super-saturated relative to ice by 10% and at -20°C by 21%' (Addison et al. 2008). The water vapour thus attempts to return to equilibrium, and so water vapour deposits onto the surface of into ice particle. As the ice crystals sink into lower layers of the cloud where temperatures are only just below freezing, they have a tendency to stick together to form snowflakes. This is aided by the supercooled droplets of water that behave as adhesives.

The micro-physical processes exposed by the Bergeron-Findeisen theory are furthermore self-perpetuating. Deposition that creates a freezing nuclei capable of inducing further droplet clustering. In such a manner, tiny ice crystals grow rapidly by deposition from vapour. Their hexagonal, dendritic forms are different however, revealing the wide range of temperatures at which they were created. Crystals formed readily aggregate upon collision due to these branched, dendritic shapes. The number of ice crystals also tends to increase exponentially because small splinters become detached by air currents during growth and act as fresh nuclei in their new domains. The freezing of supercooled water drops further adds to the production of ice splinters. The crystals formed readily aggregate upon collision due to their branched, dendritic shape. Eventually, the droplets formed become sufficiently large and heavy to overcome the lift provided by updraught, and thus they begin to fall through the cloud into the clear air beneath. As Barry et. al posit, 'when the fall speed of the growing ice mass exceeds the existing velocities of the air up-currents the snowflake falls, melting into a raindrop if it falls about 250m below the freezing level' (Barry & Chorley, 2010).

Climatologists are further understanding the complexity of cloud formations, positing the existence of interactions within but also between clouds. In a process of natural seeding, O'Hare describes how when ice crystals fall from high-level cirrostratus clouds - the seeder - into nimbostratus clouds - the feeder - composed of supercooled water droplets, the latter is fed to grow rapidly through the Bergeron process. Such occurrences are frequent in winter cyclonic systems and often lead to extensive and prolonged precipitation. However, whilst the Bergeron-Findeisen theory accounts for much precipitation in the middle and higher latitudes that are capable of sub-zero atmospheric conditions, it becomes less pronounced in the tropics. Cumulus clouds over tropical oceans precipitate when they are only 2000m deep and the cloud-top temperature is 5°C or more, far too high for supercooling to take effect. In similar vein, midlatitude summer precipitation occurs without the presence of a subfreezing layer. This climatological anomaly suggests that there are further mechanisms affecting the formation of precipitation droplets.

Theories of raindrop collision and coalescence exist as corollary and extension to the condensational processes defined by the Bergeron-Findeisen theory. By accounting for the deep convectional currents and turbulence within rising thermals, coalescence theory underlies much of the rapid creation of clouds. As Barry et. al write, 'it was originally thought that cloud particle collisions due to atmospheric turbulence would cause a significant proportion to coalesce.' However, it was soon observed that particles just as easily break up if subject to violent collisions. Langmuir thus offered a variation on this idea, suggesting that falling drops tend to have terminal velocities directly related to their diameters, such that the large drops would overtake and absorb small droplets (Barry & Chorley, 2010). Climate observation in the maritime tropics suggests that 'although coalescence is initially slow, droplets reach 100-200um radius in 50 minutes. This rapid onset of growth is facilitated by the presence of giant nuclei in the form of salt particles, suspended in a high liquid water content atmosphere. Conversely, continental air tends to contain many small nuclei and less liquid water. These differences underlie the rapid onset of showers in maritime clouds yet the more incremental precipitation in continental landmasses. Tropical cumulus clouds of the former grouping are forced into precipitation by collisions within its unstable, rising and sinking interior. Turbulence and strong updraughts as such serve to encourage collision in the early stages of cloud formation.

By dissecting hail, this growth mechanism is rendered most visibly acute within its layered, depositional structure. The coalescence process thus encourages more rapid growth than simple condensation because it increases the interactions between differentially heated cloud layers , allowing for vapour and condensation nuclei to coagulate and form precipitation droplets. As Addison et. al conclude, 'the rate at which vapour is converted into water droplets and precipitation depends upon three main factors: the rate of coalescence and ice crystal growth; the cloud thickness; and the strength of updraughts in the cloud.' (Addison et. al, 2008).

The deep interactions and symbioses expressed between macro-mechanical processes of moving air and micro-chemical processes of condensation underlie the creations of different precipitation droplets. Rain, by in large, is a consequence of droplets overcoming the lift provided by updraughts and subsequently falling through the cloud into the clear air beneath. As the volume of water in these large droplets is large relative to their surface area, little evaporation occurs in the non-saturated air below the cloud. Rather, the droplet gradually morphs from a solid ice state into a liquid one before reaching the ground. This process of state change is however non-existent when, in cold conditions, the underlying air at or near the surface is near 0°C. In such occasions snowfall occurs. Similarly sleet occurs in cold atmospheric conditions in state strung between snow and rain. This is the case when falling, the droplets encounter a freezing layer close to the surface that reforms the liquid droplets into ice pellets. These intimate surface-atmosphere interactions are further revealed in the case of ripe. Ripe forms as clear crystalline, granular ice when supercooled fog encounters a vertical structure such as trees or cable. It commonly occurs in cold, maritime climates and on mid-latitude mountains in winter. Similarly dew forms when condensation droplets on the ground surface or grass are deposited when the surface temperature is below the air's dew-point temperature.

Hail however engages in the most volatile formation, created very much at the interface between mechanical and chemical processes. Hail pellets originate when the Bergeron process operates in a cloud with a small liquid water content. Ice pellet growth predominates through the deposition of water vapour. The pellet's internal structure is however not uniform, a consequence of its turbulent vertical and horizontal motions throughout the cloud. As such hailstones are roughly concentric accretions of clear and opaque ice. The embryo is a raindrop carried aloft in an updraft and subsequently frozen. Thunderstorms continually recycle hailstones before precipitating them out. Indeed, the stone is involved in complex movements within the cloud, being swept up to the higher colder parts of the cloud several times. When this occurs, any moisture condensing on the stone freezes instantly, including any trapped air, producing opaque ice. Similarly the clear ice (glaze) within hailstones is a representative of its formational history. The wet surface layer of glaze develops as the result of rapid collection of supercooled drops in parts of the cloud with large liquid water contents which has subsequently frozen (Barry & Chorley, 2009). These interactions - of turbulent, violently fluctuating up-currents and the internal, chemical properties of water and ice - are deeply embedded in the interval between the moment of evaporation and precipitation at the Earth's surface.

Climatologists identify three main types of precipitation involving these droplet types - convective, cyclonic and orographic. Convective precipitation is strongly associated with towering cumulus and cumulonimbus clouds. These cells of activity are then often subcategorised according to their degree of spatial organisation. Of the primary subcategory, the scattered convective cells - spontaneous rising of moist air is facilitated by strong surface heating. These thermals thus often occur in the warm and humid regions of the Earth in summer months when solar heating creates the steep lapse rates, characteristic of instability and convection. The buoyancy and convective instability of the air is further reproduced by low temperatures in the upper troposphere created by the long-wave radiational heat loss of cold, cloudless summer nights. The second subcategory refers to showers of rain, snow and soft pellets that form when cold, moist, unstable air passes over a warm surface. Such an occurrence is common in midlatitude depressions, the convective cells being located parallel to the surface cold front in the warm sector. The third subcategory of convectional precipitation manifests itself in tropical cyclones and deep cumulonimbus clouds wherein the individual thermal cells become organised around a center in spiraling bands. Convergent or cyclonic precipitation defines a different mode of uplift. Rather than being determined by vertical strong surface heating, cyclonic precipitation is created by the ascent of air through horizontal convergence of airstreams in areas of low pressure. Such is the case in the hurricane-spurning, equatorial low pressure regions wherein troughs create large cyclonic precipitation through airstream convergence in the tropical easterlies. The final category of precipitation is regarded as orographic in nature. Orographic precipitation originates where air meets an extensive barrier and its subsequently forced to rise. The complexities of orographic precipitation are explored by Barry et. al. They suggest that barriers, such as mountain belts, may produce several different effects, depending on their alignment and size. These include: forced ascent on a smooth mountain slope, producing adiabatic cooling, condensation and precipitation; triggering of conditional instability by blocking of the airflow and upstream lifting; triggering of convection by diurnal heating of slopes and up-slope winds; precipitation from low level cloud over the mountains by 'seeding' of ice crystals from an upper-level feeder cloud; an increased frontal precipitation by retarding the movement of cyclonic systems and fronts (Barry & Chorley, 2009).

Such orographic precipitation is furthermore a self-perpetuating process, underlined by dynamic vertical sways from atmospheric stability to instability. If surface air is moist but atmospheric air above is drier, as the parcel of air rises, the rates of cooling between the top and bottom of the layer will be different. As such the upper cools more quickly along the dry adiabatic lapse rate that the lower, saturated air following the saturated adiabatic lapse rate. Th upper part thus becomes less stable more rapidly, facilitating strong vertical, convective air motions. As Addison et. al posit, this situation is known as convective or potential instability (Addison et. al, 2008). The conditions that produce vertical atmospheric movement are thus facilitated by the interdependent mechanisms of convectional, convergent and orographic uplift. It is through such movement and vertical energy transfer, that surface evaporation interacts with atmospheric condensation. Precipitation permanently recycles these interactions, and in doing so reproduces the dynamic Earth climate.

Thunderstorms are the most visible manifestations of the processes occurring during the interval between evaporation and precipitation. Indeed, Barry et. al characterise the mid-latitude thunderstorm as 'the most spectacular example of moisture changes and associated energy releases in the atmosphere. Its development is thus underpinned by a dual recognition of the fluid, thermodynamic processes and microphysical processes inherent in the atmosphere. The violent upward and downwards movements of air within the thunderstorm, are both the principal ingredients and motivating machinery of its development. These mechanisms of vertical energy transfer originate, as Brugge posits, from the three precipitation types outlined above; due to rising cells of excessively heated moist air in an unstable air mass; through the triggering of conditional instability by uplift over mountains; or through mesoscale circulations along convergence lines. As condensation begins to form cloud droplets, latent heat is released and the initial upward impetus of the air parcel is augmented by an expansion and a decrease in density until the whole mass becomes completely out of thermal equilibrium with the surrounding air. This constant release of latent heat thus continuously injects fresh supplies of energy that serve to energise and accelerate the updraughts into behaving buoyantly. Such conditional instability has the tendency to persist in the atmosphere, a facet observable in the large geographical trajectories of many tropical storms. Positive feedbacks which serve to further destabilise the atmosphere occurs when downdraughts, created by the friction of falling droplets, reaches the surface and spreads out as a cold pool. This process forms a vital part of the storm structure because the cold pool boundary acts as a convergence zone at which boundary-layer air is lifted. It thus provides a region - known as the 'gust zone - in which the atmosphere is inclined towards new turbulent, updraught growths. As these downdraughts gather, cold air may eventually spread out below the thunder cell in a wedge. Gradually, as the parcel's moisture is expended, the supply of latent heat diminishes, the downdraughts usurp the warm updraughts, and the cell dissipates, returning the atmosphere to relative stability. Earth's outer fluid spheres operates as an interconnected system of moving air and water that fundamentally creates and controls the hydrological system and global climate. The processes of evaporation and precipitation are deeply embedded within this system of dynamic moisture, energy and heat movement and recycling. This paper has fleshed out the macro-micro processes involved in the creation of such droplets from the skies. Cloud formation is facilitated dually by the mechanical movements of moisture and the chameleonic, chemical behaviours of water particles. The interval between evaporation and precipitation is thus identified as one of chaotic yet structured anarchy.

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